

Riso National Laboratory, Denmark
METEOROLOGY AND ATMOSPHERIC DISPERSION IN A COASTAL AREA
: Proceedings from a EURASAP Conference held at Riso
National Laboratory, Denmark, October 25-27th, 1988
1988.

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Atmospheric Dispersion in a Coastal Area

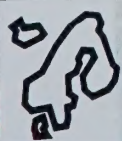
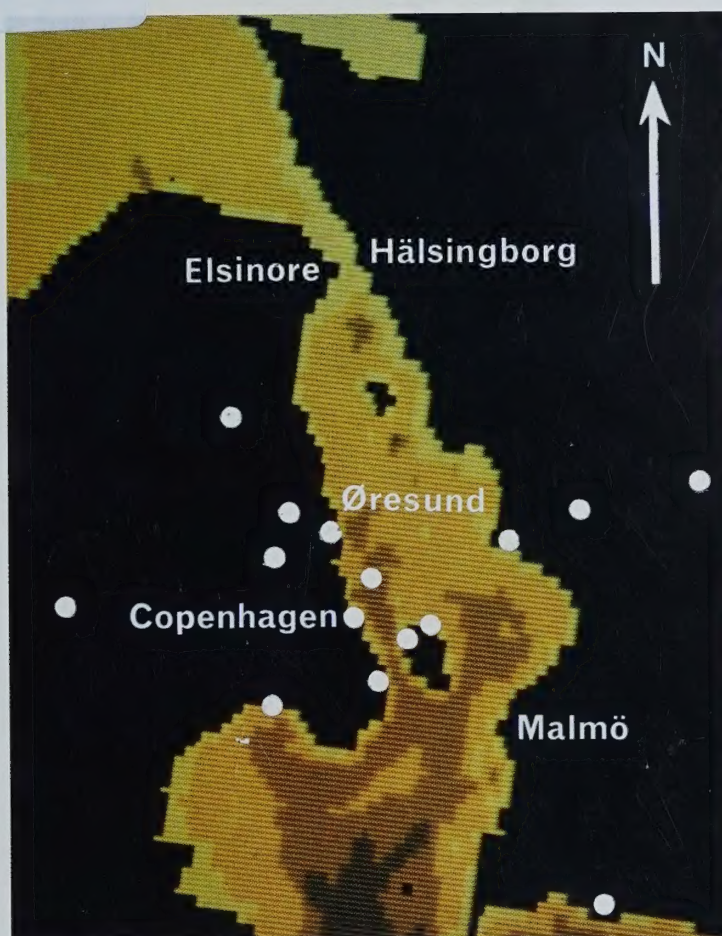
Proceedings from a EURASAP conference

held at

Risø National Laboratory, Denmark

October 25-27th. 1988

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Meteorology and Atmospheric Dispersion in a Coastal Area

Proceedings from
EURASAP conference
The ØRESUND EXPERIMENT, Workshop III

held at

Risø National Laboratory, Denmark
October 25-27, 1988

Edited by Sven-Erik Gryning

Meteorology and Atmospheric Dispersion in a Coastal Area

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Preface

The Øresund experiment was carried out in 1984 in order to investigate the modifications in the meteorology and atmospheric dispersion processes over a land-water-land area, with warm land and cold water surfaces.

The first Øresund-workshop, held in Lyngby 8–9 October 1986, dealt primarily with the presentation of the measurements.

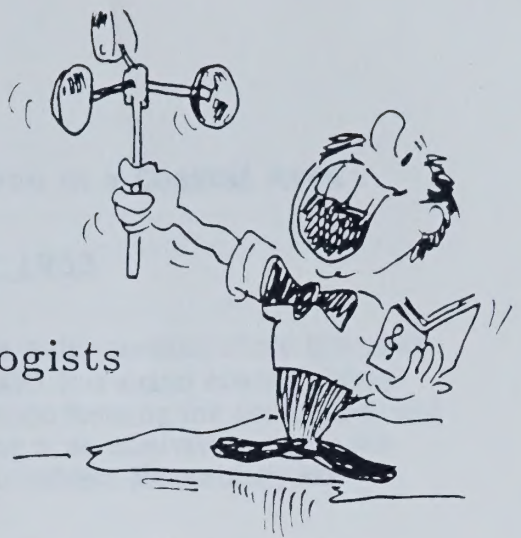
The second Øresund-Workshop, Uppsala 13–14 October 1987, was directed towards scientists using the Øresund data-bank. During the workshop it became clear that our understanding of the mechanisms that control the atmospheric dispersion even in a relatively simple area as encountered in the Øresund-experiment is not very good.

In order to present the Øresund-experiment and the results of the data analysis to a broader audience than in the foregoing workshops, and to give a summary of the state of art on meteorology and atmospheric dispersion in a coastal environment, the third Øresund-workshop is carried out as an EURASAP-conference.

The conference was arranged by a committee with the following members: Helen ApSimon, England; Mari-Mai Lagus, Norway; A.S. Smedman, Sweden; Sylvain Joffre, Finland; Bjarne Sivertsen, Norway; and myself. I thank the committee for easy co-operation and especially Mari-Mai Lagus for her considerable contribution in the arrangement. Also I thank Ulla Riis for taking care of the practical details.

Risø National Laboratory
30th of November 1988.

Sven-Erik Gryning



Song to the coastal meteorologists

Melody: Daisy or "A bicycle built for two"

1. Coastal problems
mesoscale roughness lengths
We are gathered
to solve inversion strengths
the unstable surface layer
is building up on you
the database
is "Øresund"
for a flow model built for two.

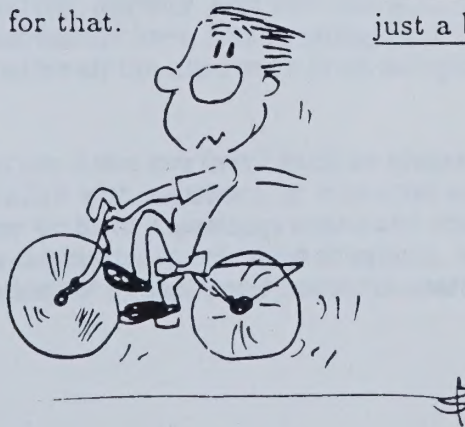
2. Wind and tracers
flow over water/land
sodars, lasers
hundreds of samplers ran.
And now is the time for knowing
what really was the clue
did Øresound
put any bound
on the flow model built for you.

3. Sven-E Gryning
made up a conference
coastal data
made it a difference
if wind was obtained in Denmark
or stress in Barsebäck
We all knew
that a common hat
was a flow model built for that.

4. Speakers language:
modelling complex case
program coding
eddy "diffusivity"
momentum and conservation
of mass and fantasy
the turbulence
the variance
in a flow model built for two

5. Further wordings:
friction velocity,
three dimensions,
sea breeze, viscosity,
the mountains, the penetrations,
the upslope circulations
it bothers you
if wind just blow
in a flow model built for two.

6. Research people
came to participate
z-nought, length scales
made matter dissipate
did anyone search solutions
for problems you never knew?
We give to you
a Danish clue
just a bicycle built for two!



Meteorology and Atmospheric Dispersion in a Coastal Area

Riso, Denmark 25-27 October 1988

Good morning. I am delighted to see you all here at this meeting of our European Association for the Science of Air Pollution (EURASAP). It is almost exactly 3 years ago that 10 scientists from all over Europe met to discuss forming the Association, and this is our 10th meeting-so this occasion is something of an anniversary. Like our other meetings this one is concentrated on a specific subject-Meteorology and Atmospheric Dispersion in a Coastal Area.

Europe has a very long coast line, and most of the European countries own some part of it; those that do not often have large lakes introducing similar effects on atmospheric behaviour. The coastal margins tend to be quite densely populated, and many industrial and power plants, which are relatively important sources of atmospheric pollutants, are sited near the sea. Thus in the UK for instance our nuclear power plants are situated on coastal or estuarine sites. In such circumstances the behaviour of the release can be radically different from that over a uniform flat surface, and it is important to appreciate how changes in surface characteristics affect the flow.

The first person from Denmark that I ever heard about was King Canute, or King Knut, who was the Dane who became King of England in the year 1016, in the same century as the Cathedral was first built in Roskilde. Now King Canute, being a wise Dane, understood the behaviour of the sea very well, and that even he as a King could not alter it. Moreover he demonstrated this to the foolish English who suggested that he could stop the tides, by an experiment at the sea shore, getting his feet wet inevitably in the process.

Now, in the twentieth century, Denmark and her Scandinavian neighbours have taken the initiative in mounting an experimental campaign on the sea. This time the aim has been to study the air flow across the Oresund Strait from Sweden to Denmark, and how the change in surface affects transport and dispersal of a pollutant in that flow. This has been a very major experiment involving coordination of scientists and experimental resources from several countries. There have already been workshops with very lively debates about some of the intriguing results obtained, and I look forward to seeing some of these questions resolved over the next three days I hope.

This experiment has also provided the initiative for this meeting, to bring together all those interested in meteorology and atmospheric dispersion in a coastal area, to pool our ideas and present our own work for discussion, and I am sure we have an interesting time ahead.

I should like to thank our organisers in advance for making this possible- Mari Mai Lagus and Bjarne Sivertsen from Norway, Ann Sofi Smedman from Sweden, Sylvain Joffre from Finland, and finally Sven Erik Gryning from Riso (who I am glad to tell you, has not been deterred by all the hard work from accepting a position on the EURASAP Committee).

On behalf of the Association can I also say that I shall be pleased to discuss any ideas or suggestions about EURASAP with members, or to recruit any of those present who are not members. Next year we have 2 meetings scheduled which may be of interest to you- one in Italy on dispersion in low wind situations, and the other on background aerosol concentrations at a remote and beautiful coastal site on the west coast of Ireland.

Returning to this meeting the programme is organised in a very logical manner, starting today with the meteorological aspects of flow over non-uniform terrain and coastal areas. I am glad to see a wide geographical representation in the contributions on this topic. Tomorrow we can hear about the Oresund experiment and the results obtained. Finally we can see how well we understand these complex flows by looking at the capabilities of models to simulate the processes involved.

I wish you all an enjoyable and constructive meeting.

Meteorology in Coastal Areas

The first of these is the fact that the
the second is the fact that the
the third is the fact that the
the fourth is the fact that the
the fifth is the fact that the

The sixth is the fact that the

The seventh is the fact that the

Wind and Temperature Structure of the Atmospheric Boundary Layer in a Coastal versus Arid Site

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Israel Electric Corp. Ltd., P.O.B. 10 , Haifa , Israel

Uri Dayan

Israel Atomic Energy Commission , Yavne 70600 , Israel

1.0 Introduction

Evaluating the dispersion of atmospheric pollutants has become a necessary feature of modern industrial activity . A wide range of techniques for performing these calculations has been devised , taking into account many features of the earth surface and especially meteorological fields in the lower atmosphere . The distance from sea shore plays an important role on the dispersion mechanism. It is reasonable therefore to expect different wind and temperature profiles for a inland site where there is no "Internal Boundary Layer" (IBL), compared to a site in the presence of a coastal IBL . The IBL is related to the layer of air below the interface formed between air whose turbulence is determined by the water surface and air whose turbulence is determined by the land surface .

Suggestions have often been made that direct measurements of turbulence parameters at the emission site should be used in pollution dispersion estimates rather than off-site data since differences in meteorological data sites could lead to unrepresentative results (Hanna et al. 1977, Irwin 1981 ,Kahl and Samson 1988 , Draxler 1988 and Dayan et al. 1988). Furthermore, simultaneous meteorological data in several sites is needed in order to test the models (Hogstrom 1987). In this view the concurrent hourly measurements 24 hour a day in two towers within 130 km , as presented here , may contribute to the understanding of mesoscale and regional effects .

The aim of this paper is to present some selected meteorological characteristics , based on simultaneous tower measurements , mainly wind and temperature . The dependence of this parameters on atmospheric stability in coastal versus inland site in Israel is investigated .

2.0 Site and instrument description

The Ramat-Hasharon (RH) tower is located at the central coast of Israel, 5 km from the seashore (Fig. 1) at an elevation of 30 m above sea level within a semi-urban area. Temperature and wind are measured at two levels: 10 and 65 m above ground.

The Shivta (SH) tower is located 60 km from the seashore, in the Central Negev (340 m above sea level) with a terrain

relief of 150 m within a radius of 10 km. The distance between the sites is about 130 km.

Instrumentation and data collection method are based on the US NRC Regulatory Guide 1.23 (1972). Wind, temperature and other parameters are measured at three levels: 10, 60 and 90 m above ground. All sensors are scanned each second. These values are averaged hourly and logged by a data logger. The hourly data for June-August 1987 for two sites have been chosen as the basis for the following presentation.

3.0 Consideration of the wind profile at RH and SH

Directional wind shear

The primary influence of the Coriolis effect on horizontal air flow is that it brings about a turning of the direction of the mean horizontal wind with height. The differences in wind shear between the coastal (RH) and the inland (SH) sites were calculated based on wind direction differences between 60 and 10 m elevation as shown in Fig. 2.

In both sites a minimal wind shear exists during daytime, for which a greatest lapse rate is observed and its maximum value is at early morning when the air is most stable. A similar behavior is shown in Fig. 3, where wind shear is related to atmospheric stability classification calculated by the AEC lapse rate method. The explanations for high wind shear is found in the suppression of vertical eddies during stable atmosphere. The coupling between upper (60 m) and lower layer (10 m), or the eddy viscosity, is smaller and momentum is not appreciably transported downward. In spite of the general resemblance the wind shear in the coastal site (RH) is larger and reaching its maximum value of up to 50° during very stable conditions (category-F) whereas the maximum directional variation in the continental site (SH) is about 20° for category E. The reason for higher wind shear exhibited in RH is due to a pronounced southerly component in the surface layer as a result of a more developed breeze effect.

Velocity wind profile

The wind velocity profile is widely represented by an empirical power law formula:

$$V_1/V_2 = (Z_1/Z_2)^P$$

where V_1 and V_2 are velocities at Z_1 and Z_2 levels respectively. P is an exponent that depends on surface roughness and atmospheric stability. For engineering applications $P=1/7$ is used, while for most dispersion models, to account for variation of speed from anemometer to emission height, specific exponent values, stability dependent, were developed.

In Fig. 4 the exponent P is given versus stability classes, based on lapse rate at 10-60 m layer, at both sites compared to exponent estimated by Irwin (1979) for urban and a rural

areas. The RH site exponent values are in good agreement, except for moderate to very stable categories, with Irwin (1979) and Touma (1977) and with contradiction to the findings of Hanafusa et al. (1986). Exponent values obtained at SH site are lower than these observed in the coastal site. At both sites P values for unstable categories (A, B and C) were found to be constant, fitting Irwin's results.

A suggested explanation for differences in P's are the unsimilar roughness conditions between the coastal and the inland sites. Results for F and G categories are of a limited value due to rather small population sampled in our data record.

4.0 Structure of the temperature profile at RH and SH

One of the most important meteorological parameters affecting the dispersion of pollutants is the vertical thermal structure of the shallow atmosphere. Hourly averaged temperature lapse rate for summer from both towers were derived and plotted on Fig. 5. The main feature found is the discrepancy of lapse rate at noon, between both sites, and is explained by the time lag of the sea breeze penetration to the inland site.

5.0 Summary and conclusions

In this paper the wind and temperature structure of the atmospheric boundary layer up to 60 m in a coastal versus arid site in Israel for the summer 1987 was evaluated. The main conclusions are as follows:

- a) During noon time, wind shear in the first 50 m layer above ground is marginal for both sites. At night, significant wind shear values up to about 30° are observed. This finding might be important for air pollution estimation, especially for low emission sources.
In unstable to neutral atmosphere, the observed wind shear, in both sites, does not exceed the value of 5° . However, in very stable atmospheric conditions, such as F category, it may reach up to 50° in average, especially in the coastal region of the country.
- b) A reasonable agreement was found for the power law exponent (P) calculated for the coastal site and Irwin's values mainly for A through E categories. In the inland site, much lower values were observed as compared to Irwin's values for rural areas. The sample for F and G categories appeared to be insufficient in order to reach a firm conclusion.
- c) The observed difference in lapse rate between the coastal and the more remote and arid site is governed by timing of the sea breeze inland penetration.

In summary, a strong spatial correlation was found between both coastal and arid site in cases of unstable atmospheric conditions. The weaker correlation observed for stable conditions can be explained by stronger local micro-meteorological influences.

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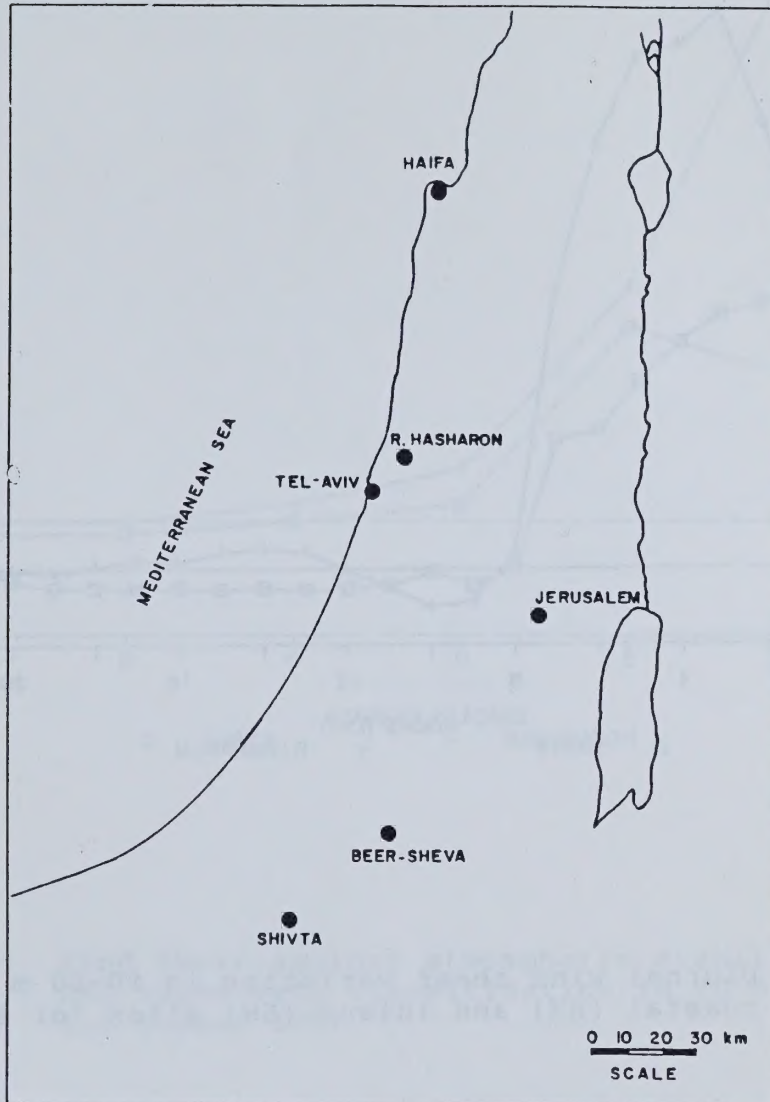


Figure 1: Location of the meteorological towers Shvta and Ramat-Hasharon

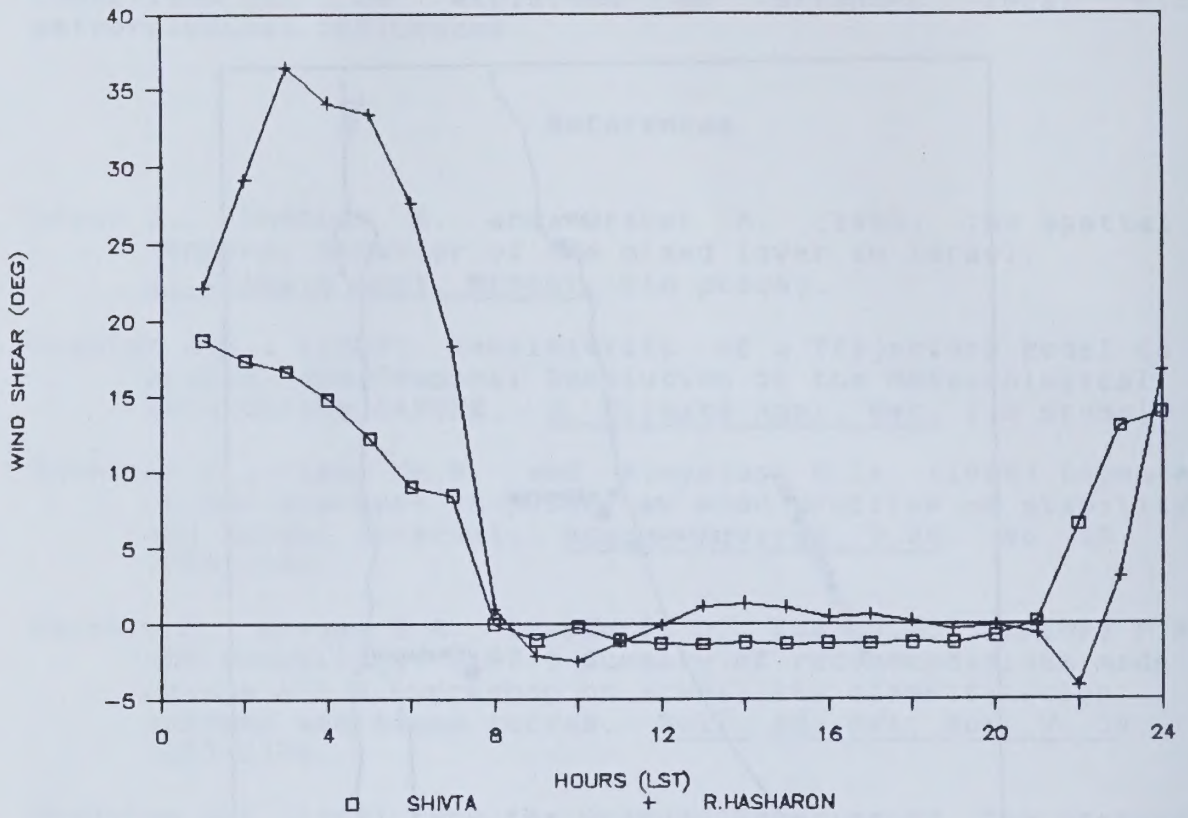


Figure 2: Diurnal wind shear variation in 10-60 m layer at a coastal (RH) and inland (SH) sites for summer 1987.

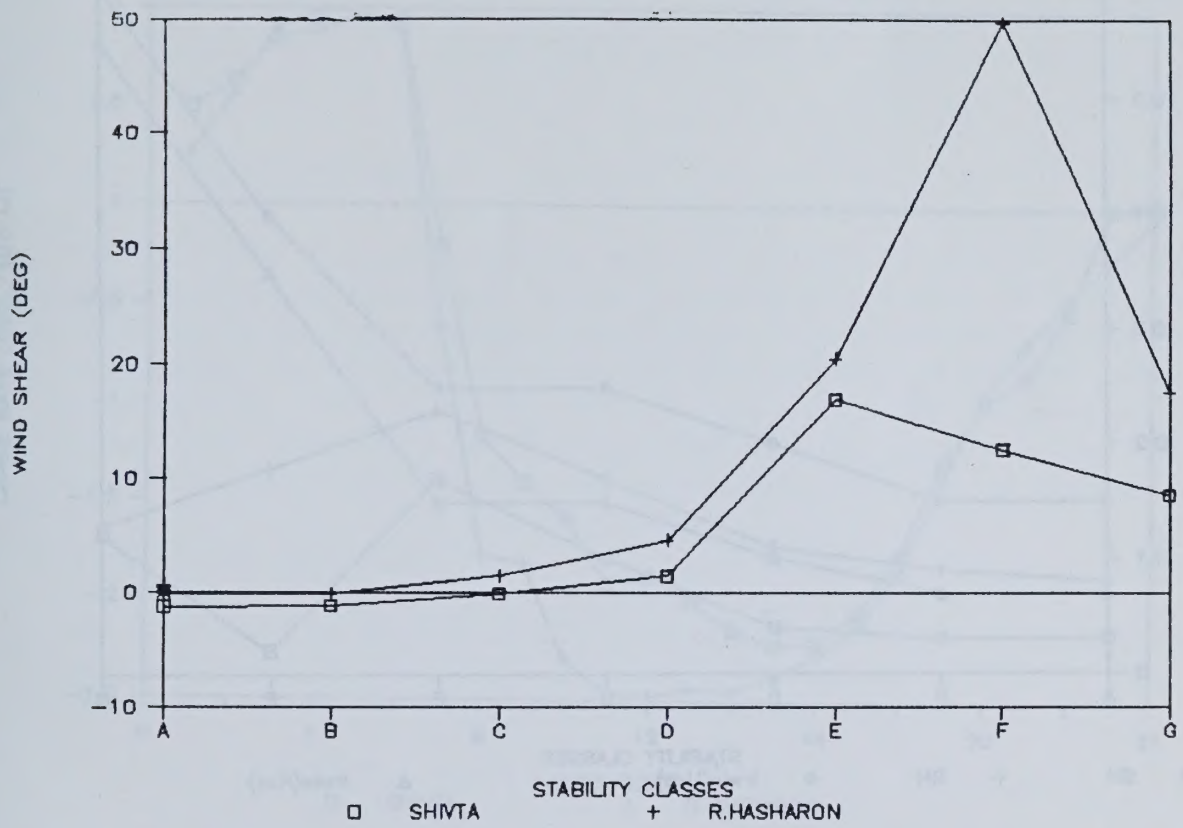


Figure 3: Wind Shear against atmospheric stability based on AEC lapse rate classification in 10-60 m layer for summer 1987.

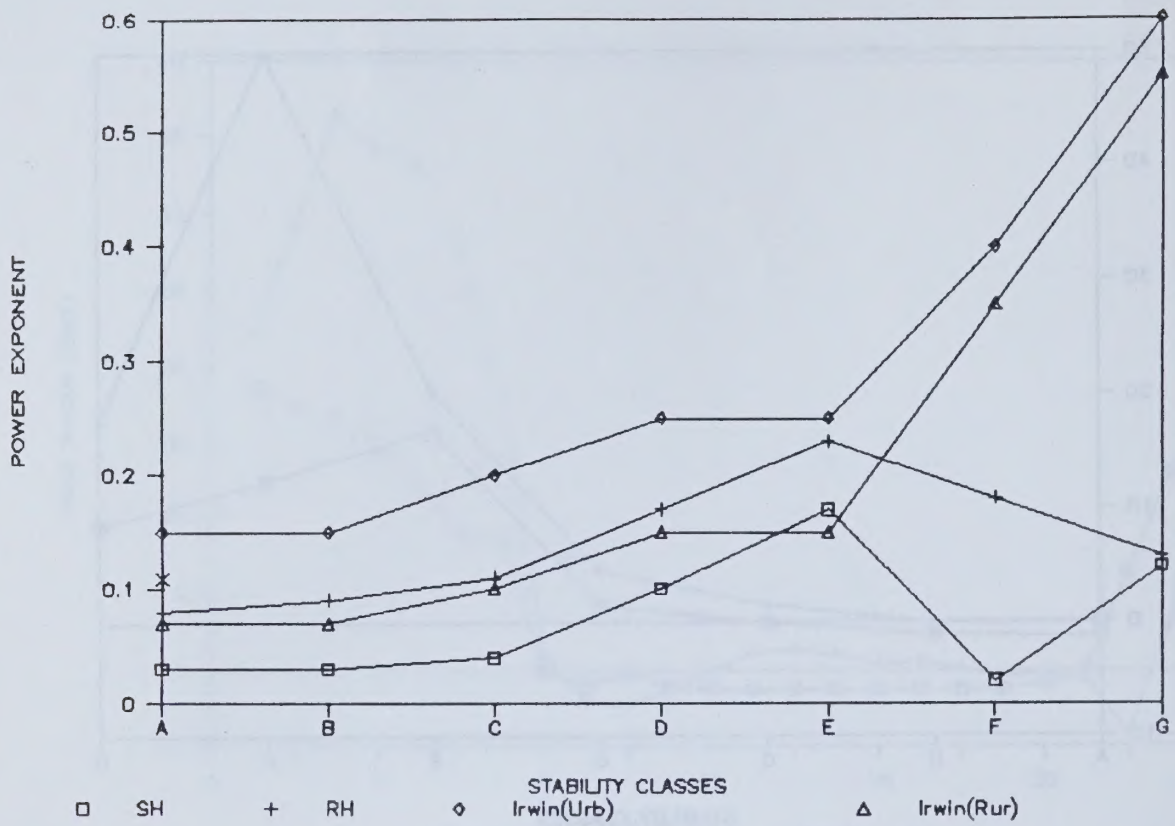


Figure 4: Power law exponent variation with atmospheric stability based on AEC lapse rate scheme in 10-60 m layer for summer 1987

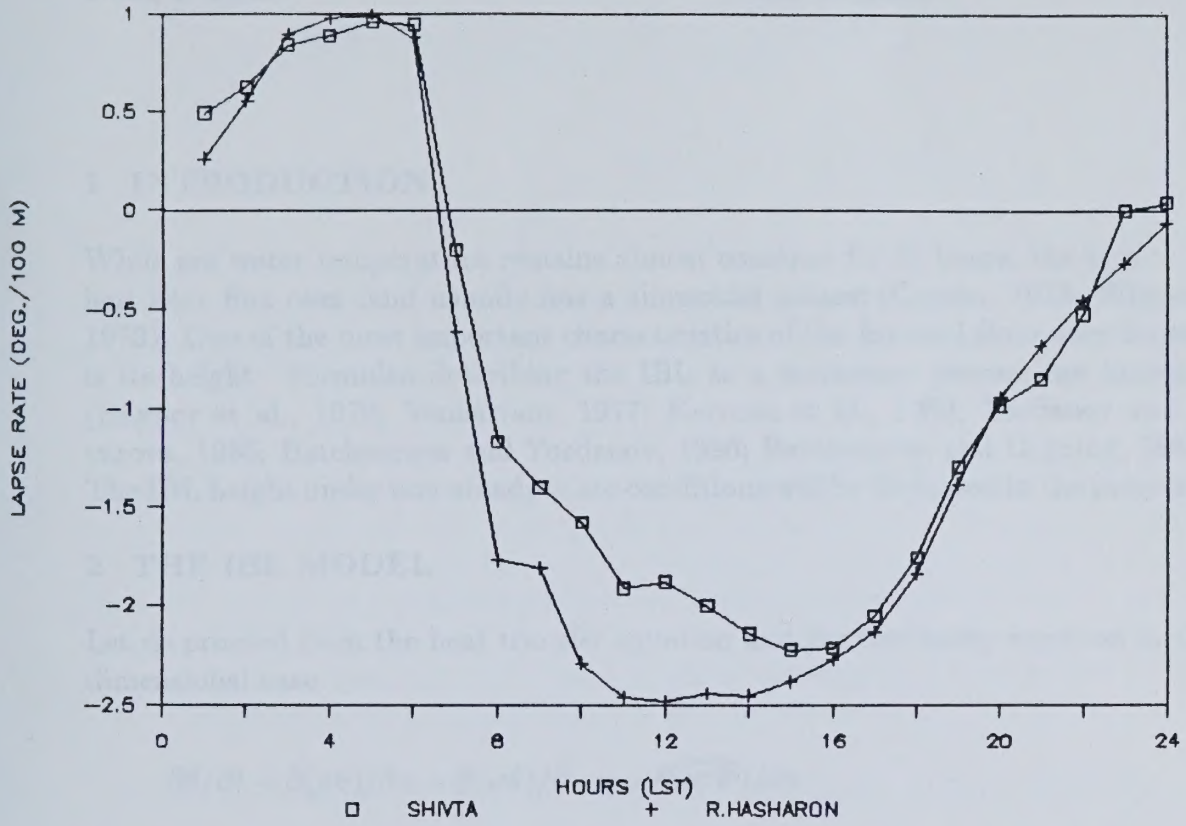


Figure 5: Diurnal lapse rate variation in 10-60 m layer at a coastal (RH) and inland (SH) sites for summer 1987.

NON-STATIONARY INTERNAL BOUNDARY LAYER

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1 INTRODUCTION

While sea water temperature remains almost constant for 24 hours, the vertical turbulent heat flux over land usually has a sinusoidal nature (Carson, 1973; Wippermann, 1973). One of the most important characteristics of the Internal Boundary Layer (IBL) is its height. Formulae describing the IBL as a stationary process are known so far (Raynor et al., 1979; Venkatram, 1977; Kerman et al., 1982; Yordanov and Batchvarova, 1985; Batchvarova and Yordanov, 1986; Batchvarova and Gryning, 1988; etc). The IBL height under non-steady-state conditions will be discussed in the present paper.

2 THE IBL MODEL

Let us proceed from the heat transfer equation and the continuity equation in the two-dimensional case

$$\partial\theta/\partial t + \partial(u\theta)/\partial x + \partial(w\theta)/\partial z = -\partial(\overline{w'\theta'})/\partial z \quad , \quad (1)$$

$$\partial u/\partial x + \partial w/\partial z = 0 \quad , \quad (2)$$

where θ is the potential temperature; u, w the wind velocity components along x - and z -axes, respectively, Fig. 1, depending on time t only, and $(\overline{w'\theta'})$ is the vertical kinematic turbulent heat flux.

Let above the upper IBL boundary h , an air mass with a constant potential temperature gradient $\Gamma = \partial\theta/\partial z$ be positioned and Δ be the temperature difference across the inversion, Fig. 1. Then the boundary condition for θ at height h has the expression:

$$\theta_h = \Gamma h - \Delta + \text{const} \quad , \quad (3)$$

where θ_h is the potential temperature at height h .

Following integration of Eq. (1) from $z = 0$ to $z = h(x, t)$, we get:

$$(\overline{w'\theta'})_0 - (\overline{w'\theta'})_h = \frac{\partial}{\partial t} \int_0^h \theta dz - \theta_h \frac{\partial h}{\partial t} + \frac{\partial}{\partial x} \int_0^h u \theta dz - u_h \theta_h \frac{\partial h}{\partial x} + w_h \theta_h \quad , \quad (4)$$

where $(\overline{w'\theta'})_0$ and $(\overline{w'\theta'})_h$ are the vertical turbulent heat fluxes near the ground and at height h which depend on time. Substituting w_h determined from Eq. (2) and θ_h from Eq. (3) into Eq. (4) leads to:

$$\begin{aligned} (\overline{w'\theta'})_0 - (\overline{w'\theta'})_h &= \frac{\partial}{\partial t} \int_0^h (\theta - \theta_h) dz + \frac{\partial}{\partial x} \int_0^h u(\theta - \theta_h) dz \\ &+ \left(\Gamma - \frac{\partial \Delta}{\partial h} \right) \left(h \frac{\partial h}{\partial t} + \frac{\partial h}{\partial x} \int_0^h u dz \right) . \end{aligned} \quad (5)$$

In Batchvarova and Yordanov (1986) and Batchvarova and Gryning (1988), Eq. (5) was obtained in the case $\Delta = 0$. Tennekes (1973) proposed the following expression for the entrainment

$$-(\overline{w'\theta'})_h = A(\overline{w'\theta'})_0 + Bu_*^3/\beta h , \quad (6)$$

where u_* is the friction velocity, β the buoyancy parameter, $A = 0.2$ and $B = 2.5$.

Under horizontally homogeneous conditions Eq. (7) is obtained for $\partial\Delta/\partial h$:

$$\partial\Delta/\partial h + \Delta [1/h + 1/(Ah - B\kappa L)] = \Gamma \quad (7)$$

with simple approximation for the decision as follows:

$$\Delta/\Gamma h = (Ah - B\kappa L)/[(1 + 2A)h - 2B\kappa L] , \quad (8)$$

where $\kappa = 0.4$ is the von Karman constant and $L = -u_*^3/\kappa\beta(\overline{w'\theta'})_0$ is the Monin-Obukhov length, see Tennekes (1973); Gryning and Batchvarova (1988).

If $\Gamma - \partial\Delta/\partial h > 0.004 K/m$ is assumed, all terms in Eq. (5) not including $\Gamma - \partial\Delta/\partial h$ may be neglected, and from Eqs. (5) - (8) follows

$$\partial h/\partial t + (u_h - u_*R)\partial h/\partial x = (1 + 2A)(\overline{w'\theta'})_0/\Gamma h + 2Bu_*^3/\beta\Gamma h^2 , \quad (9)$$

where

$$u_h - u_*R(\mu) = (1/h) \int_0^h u dz , \quad (10)$$

u_h is the wind velocity at height h , and $\mu = h/L$ is the internal stratification parameter. If the Businger wind velocity profile is used, Gryning and Batchvarova (1988) have obtained for the dimensionless function $R(\mu)$ the expression

$$R(\mu) = [1 - (1 - 16\mu)^{3/4}] / 12\kappa\mu \quad . \quad (11)$$

At $|\mu| > 1$, Eq. (11) has the asymptotic version

$$R(\mu) = 2/3\kappa |\mu|^{1/4} \quad . \quad (12)$$

According to Zilitinkevich (1975)

$$R(\mu) = r_1\kappa^{1/3} |\mu|^{-1/3} \quad (13)$$

where $r_1 = 7.5$. For big values of $|\mu|$ this difference is due to the fact that Businger's formulation does not satisfy the similarity theory requirements. Gryning and Batchvarova (1988) suggest $R(\mu)$ from Eq. (11) to be assumed constant, introducing a characteristic IBL height value equal to 0.05 – 0.1 from the land fetch. According to this, $R = \text{constant}$ is used further on.

3 THERMAL ENTRAINMENT FACTOR DOMINATION

If only the thermal entrainment factor is considered, Eq. (9) can be presented as

$$\partial h^2 / \partial t + u(t) \partial h^2 / \partial x = 2(1 + 2A)(\overline{w'\theta'})_0 / \Gamma = a_1 f(t) \quad , \quad (14)$$

where $u(t) = u_h - u_* R = u_h(1 - C_D R)$ and $f(t) = (\overline{w'\theta'})_0$ must be integrable functions of time, $C_D = u_*/u_h$ is the drag coefficient, $a_1 = 2(1 + 2A)/\Gamma$.

Linear equations with partial derivatives of first order like Eq. (14) may be expressed as:

$$\partial y / \partial t + u(t) \partial y / \partial x = a_1 f(t) \quad , \quad y = h^2 \quad . \quad (15)$$

By substitution

$$y = v + a_1 \int_{t_0}^t f(t) dt \quad (16)$$

in Eq. (15), we obtain the homogeneous equation

$$\partial v / \partial t + u(t) \partial v / \partial x = 0 \quad . \quad (17)$$

After substituting with

$$\tau = \int_{t_0}^t u(t) dt \quad (18)$$

we get

$$\partial v / \partial \tau + \partial v / \partial x = 0 \quad . \quad (19)$$

All solutions of the differential equation, Eq. (19), are given with the expression:

$$v = w(x - \tau) \quad , \quad (20)$$

where w is any function (Courant, 1962).

Let us assume that at the initial moment $t = t_0$, the height of the IBL is a known function of x , the distance from the shore

$$h = F(x) \quad , \quad (21)$$

then $y = h^2 = F^2(x) = v$ at $t = t_0$. From Eq. (20) we get for the moment t that $v = F^2(x - \tau)$, and after applying Eqs. (16) - (18) we finally get

$$h = \left\{ F^2 \left[x - \int_{t_0}^t u(t) dt \right] + a_1 \int_{t_0}^t f(t) dt \right\}^{1/2} \quad . \quad (22)$$

A direct check shows that Eq. (22) satisfies the differential Eq. (14) and the initial condition Eq. (21).

In Yordanov and Batchvarova (1988) is shown that similar considerations are also possible in the case $\Gamma - \partial \Delta / \partial h = 0$.

4 MECHANICAL ENTRAINMENT FACTOR DOMINATION

If only the mechanical entrainment factor is taken into account, Eq. (9) will be changed as follows:

$$\partial h / \partial t + u(t) \partial h / \partial x = 2 B u_*^3 / \beta \Gamma h^2 \quad . \quad (23)$$

This equation can be transformed as:

$$\partial h^3 / \partial t + u(t) \partial h^3 / \partial x = 6 B u_*^3 / \beta \Gamma = b f(t) \quad . \quad (24)$$

Assuming $y = h^3$ we obtain

$$\partial y / \partial t + u(t) \partial y / \partial x = b f(t) \quad . \quad (25)$$

Following the procedure presented by Eqs. (16) - (21), we finally get

$$h = \left\{ F^3 \left[x - \int_{t_0}^t u(t) dt \right] + b \int_{t_0}^t f(t) dt \right\}^{1/3} \quad . \quad (26)$$

If the wind velocity does not change in the afternoon hours, then stationary conditions exist and

$$h = F(x) = \left[6 B u_*^3 / \beta \Gamma u \right]^{1/3} \quad (27)$$

is assumed.

5 COMPARISON WITH NANTICOKE EXPERIMENTAL DATA

Let us for example examine the Nanticoke experiment of 1st June 1978 as a non-steady-state process (Kerman et al., 1982; Portelli et al., 1982).

In Fig. 2 the experimental data on $(\overline{w'\theta'})_0$ at 1100, 1200, 1300 and 1400 hours LST represented by circle are approximated with the function $(\overline{w'\theta'})_0 = f(t) = 0,22 \sin \pi(2 - t/t_c)$ where $t_c = 9$ when in units of hours and $t_c = 32400$ when in units of seconds. At $t = 18$ LST, we assume that the turbulent heat flux over land is $(\overline{w'\theta'})_0 = 0$ according to other experimental data for the month of June (Carson, 1973; Wippermann, 1973). It is supposed that at $t = t_0 = 13.5$ LST stationary conditions have occurred and the maximal value of the turbulent heat flux over the land $(\overline{w'\theta'})_{0,max} = f(13.5) = 0.22 mK/s$ is reached. Then for the case of thermal entrainment only

$$h = \left[2(1 + 2A)(\overline{w'\theta'})_{0,max} / u \Gamma \right]^{1/2} x^{1/2} \quad . \quad (28)$$

The afternoon IBL height can be calculated using Eq. (22). If $u(t)$ and $f(t)$ depend on time in the same way [for example $u(t) = \text{const}$ and $f(t) = \text{const}$, or $u(t) = u_{max} \sin \pi(2 - t/t_c)$ and $f(t) = (\overline{w'\theta'})_{0,max} \sin \pi(2 - t/t_c)$], the solution in the non-stationary case coincides with the stationary one. Considering the case when $u(t)$ and $f(t)$ are different functions of time, the non-stationary solution differs from the stationary one. As an

example let $u(t) = \text{const}$ and $f(t) = (\overline{w'\theta'})_{0,max} \sin \pi(2 - t/t_c)$, then we have the non-stationary solution

$$h = \left[2(1 + 2A)(\overline{w'\theta'})_{0,max}/u \Gamma \right]^{1/2} \left[x - u(t - t_0) + u \frac{t_c}{\pi} \cos \pi \left(2 - \frac{t}{t_c} \right) \right]^{1/2}. \quad (29)$$

In the course of the experiment the averaged wind speed, potential temperature gradient, friction velocity, and the Monin-Obukhov length are $u_h = 5$ m/s, $\Gamma = 0.01$ K/m, $u_* = 0.5$ m/s, and $L = -50$ m, respectively. R is supposed to be 1.3 according to Eq. (11) and for the characteristic IBL height value of 100 m.

A comparison with the experimental data is presented in Fig. 3a where both the asymptotes Eqs. (27) and (28), and the exact solution

$$\frac{u_h - u_* R}{1 + 2A} \left\{ \frac{h^2}{2} + \frac{2B\kappa L}{1 + 2A} h + \left(\frac{2B\kappa L}{1 + 2A} \right)^2 \ln \left(-\frac{1 + 2A}{2\kappa BL} h + 1 \right) \right\} = \frac{(\overline{w'\theta'})_0}{\Gamma} x \quad (30)$$

of Eq. (9) according to Gryning and Batchvarova (1988) are shown.

The point where the IBL height according to both the asymptotes coincides, is then the distance respectively the stratification parameter value up to which is better to use the mechanical entrainment asymptote and after which is better to use the thermal entrainment asymptote, Fig. 3b. From Eqs. (27) and (28) follows

$$|\mu_1| = h_1/|L| = 3\kappa B/(1 + 2A) = 2.14$$

$$\text{at } x_1 = 4.5 \beta \Gamma u B^2 \left[\frac{\kappa h}{(1 + 2A)u_* |\mu_1|} \right]^3 \quad (31)$$

For the experiment considered $x_1 = 600$ m.

Figure 4 shows the IBL height in the afternoon hours. Stationary conditions occur for nearly two hours at midday when the turbulent heat flux may be considered also constant. The turbulent heat flux decrease in the afternoon causes a IBL height decrease according to the non-stationary thermal entrainment asymptote. So the IBL height in the non-stationary case for the moment of time t could be determined following the dynamic entrainment asymptote up to the distance where both asymptotes coincide at this moment and following the thermal entrainment asymptote thereafter. At $t = 16.5$ LST for example, the IBL height should be calculated following Eq. (27) up to $x = x_4$ and following Eq. (29) at $x > x_4$.

Following the presented type of considerations, modified decisions can be obtained. For example, Eq. (9) can be rewritten as

$$\partial h / \partial t + u(t) \partial h / \partial x = (2Bu_*^3 / \beta \Gamma h^2) [1 + (1 + 2A) |\mu_5| / 2 \kappa B] , \quad (32)$$

which leads to the appearance of a factor 1.33 in the right part of Eq. (27) and represents the mechanical entrainment modified with the influence of the thermal entrainment. In this case a characteristic IBL height at a distance $x_1/2$ is used to obtain the parameter $|\mu_5|$. The corresponding IBL height is presented in Fig. 3a and 3b too, using $1.33 h$ for the mechanical entrainment asymptote up to $x = x_1$ and $h + 30$ for the thermal entrainment asymptote after that.

6 CONCLUSION

Several asymptotic solutions have been proposed, and also a way to obtain other similar approaches has been shown in this paper. Finally, the choice of formulation to be used depends on the observed meteorological characteristics.

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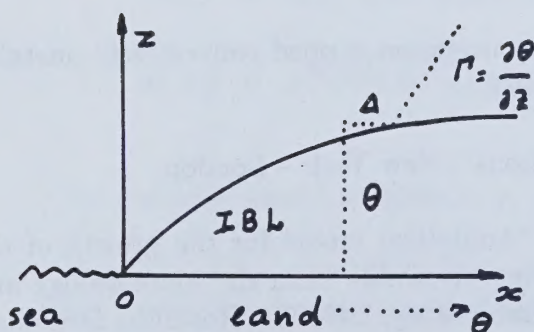


Fig. 1. IBL development scheme.

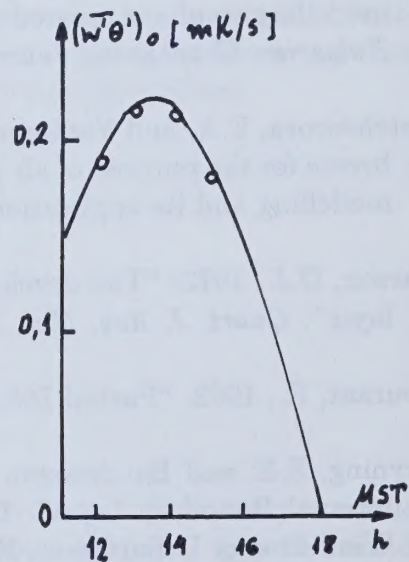


Fig. 2. The turbulent heat flux near the ground - experimental data, Nanticoke, 1 June 1978, and approximation.

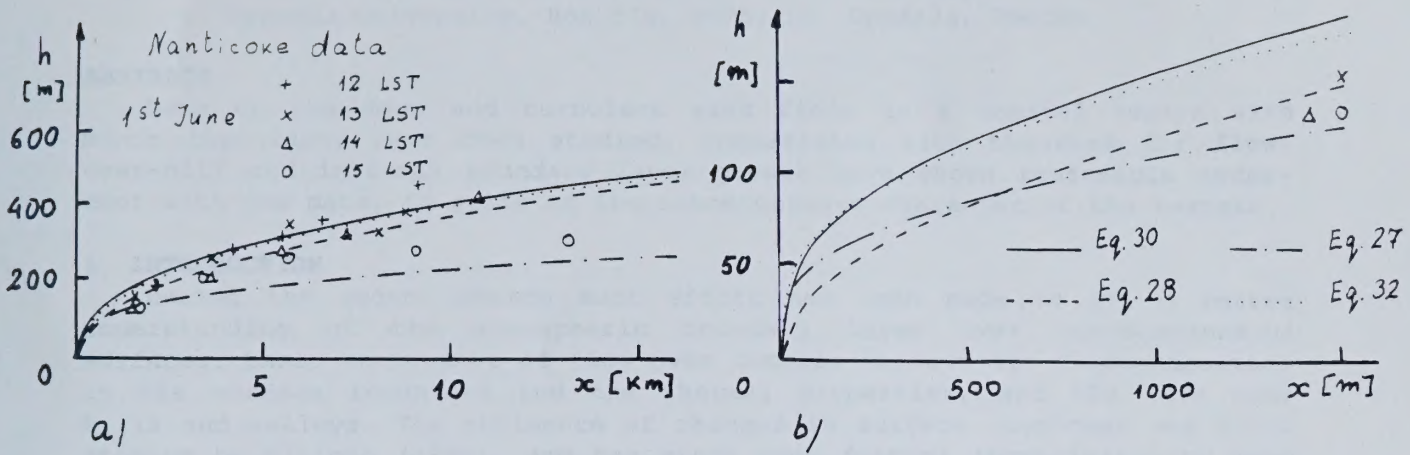


Fig.3. The IBL height – experimental data, Nanticoke 1 June 1978 and modelling results for a) all the experimental distance, b) not far from the coast.

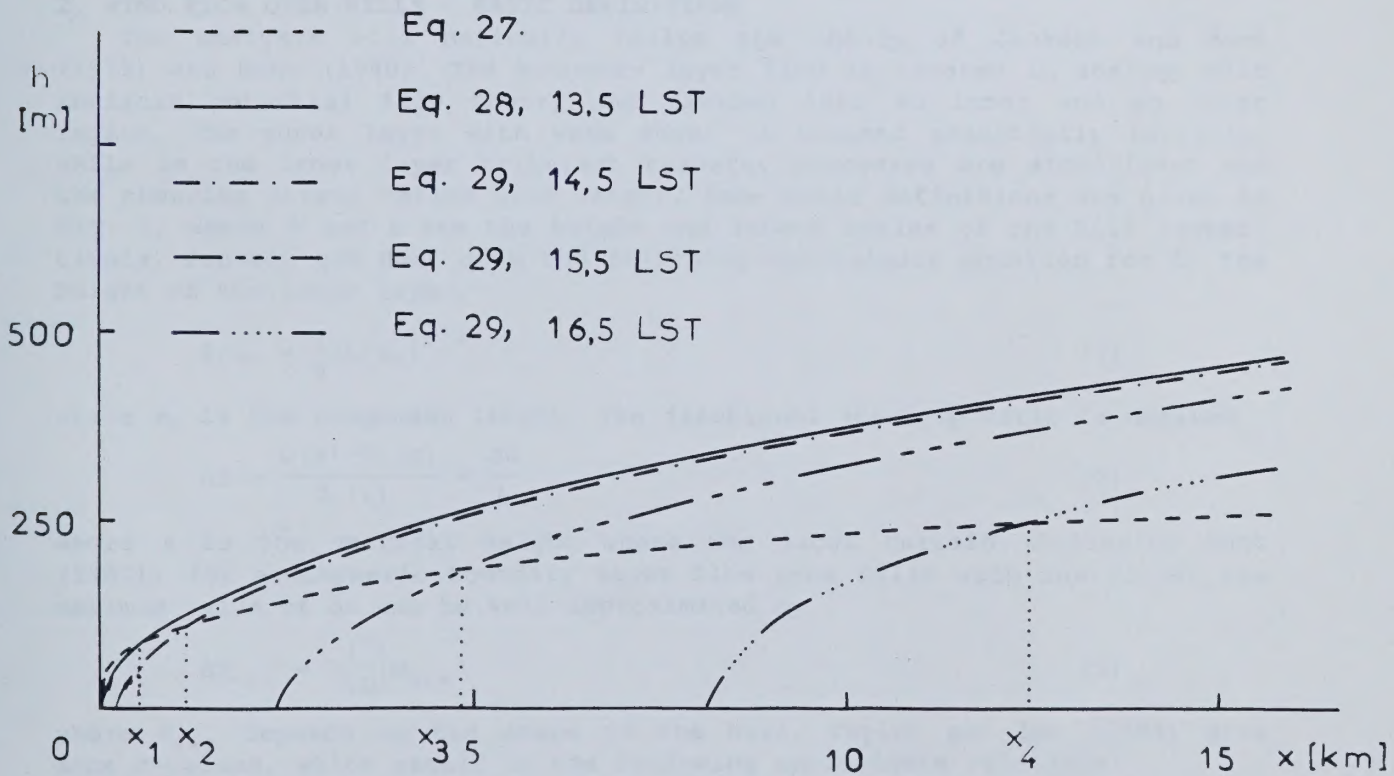


Fig 4. Non-stationary IBL.

COASTAL MODIFICATION OF ATMOSPHERIC AIR FLOW - A CASE STUDY

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ABSTRACT

Data on the mean and turbulent wind field in a coastal region with minor topography have been studied. Comparisons with theories for flow-over-hill and internal boundary layer growth have shown reasonable agreement with the data, in spite of the inhomogeneous character of the terrain.

1. INTRODUCTION

During the recent decade much effort has been made to get a better understanding of the atmospheric boundary layer over non-homogeneous surfaces. Basic components of flow over complex terrain are discontinuities in the surface roughness and the thermal properties, and the flow over hills and valleys. The influence of changes in surface roughness was first studied by Elliott (1958), and has since been further investigated by many authors - see e.g. Hunt and Simpson (1982) for a review. Regarding flow over hills a number of theoretical models has been proposed in the literature, see e.g. Jackson and Hunt (1975), Mason and Sykes (1979), Walmsley et al. (1982, 1986), and Taylor et al. (1987). The aim of the present study is to analyse some observations taken in a coastal region with minor hills. The data will be compared with simple boundary layer wind models, suitable for practical use. These models, on one hand, deal with the modification of the wind field caused by a surface discontinuity, on the other hand with the 'speed-up' observed over a hill.

2. WIND FLOW OVER HILLS - BASIC DEFINITIONS

The analysis will basically follow the theory of Jackson and Hunt (1975) and Hunt (1980). The boundary layer flow is treated in analogy with inviscid potential flow theory and divided into an inner and an outer region. The outer layer with weak shear is assumed essentially inviscid, while in the inner layer turbulent transfer processes are significant and the shearing stress varies with height. Some basic definitions are given in Fig. 1, where h and L are the height and length scales of the hill respectively. Jackson and Hunt give the following approximate equation for ℓ , the height of the inner layer,

$$\ell/z_0 = \frac{1}{8}(L/z_0)^{0.9} \quad (1)$$

where z_0 is the roughness length. The fractional speed-up ratio is defined

$$\Delta S = \frac{U(z) - U_0(z)}{U_0(z)} = \frac{\Delta U}{U_0} \quad (2)$$

where z is the vertical height above the local terrain. Following Hunt (1980), for atmospheric boundary layer flow over hills with low slope, the maximum value of ΔS can be well approximated by

$$\Delta S_{\max} \approx 2 \left(\frac{h}{L} \right) \sigma_{\max} \quad (3)$$

where $\sigma_{\max}$ depends on the shape of the hill. Taylor and Lee (1984) give some σ -values, which result in the following approximate relations:

$$\Delta S_{\max} \approx \begin{cases} 2 h/L & \text{for 2-dim. ridges} \\ 0.8 h/L & \text{for 2-dim. escarpments} \\ 1.6 h/L & \text{for 3-dim. axisymmetric hills} \end{cases} \quad (4)$$

These estimates should only be used for h/L values less than about 0.5.

Also the atmospheric turbulence is affected by the presence of hills.

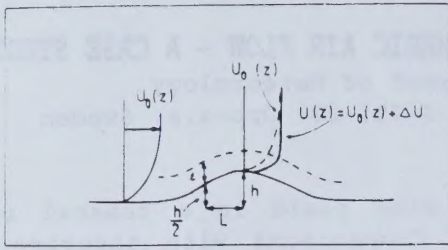


Fig. 1. Basic definitions.

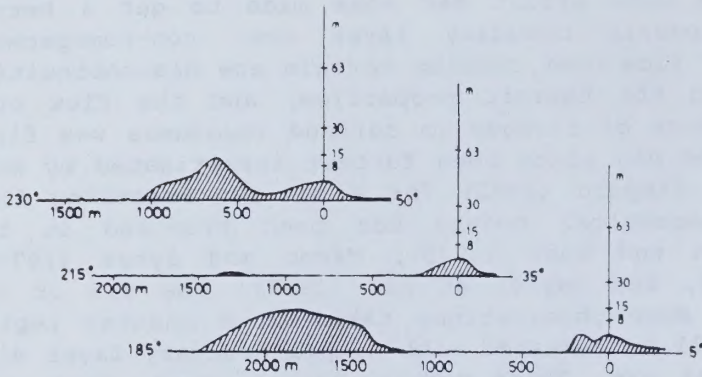


Fig. 3. Surface profiles representative of the three sectors.

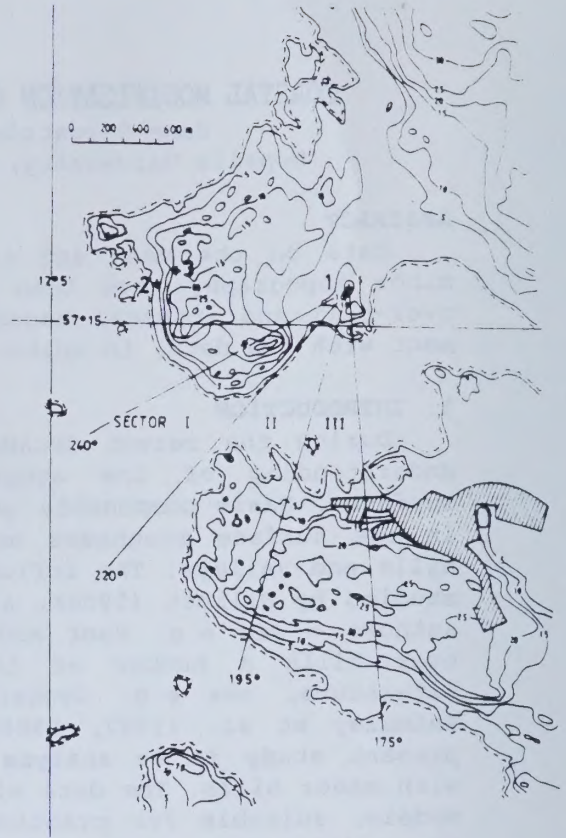


Fig. 2. Map of the experimental area.

Some estimates are presented by Hunt (1980) and Britter et al. (1981). Close to the surface, where the travel time over the hill is large compared to the Lagrangian time scale, the turbulence is close to equilibrium and Britter et al. (1981) found that for $z \ll \ell$

$$\frac{\Delta \sigma_u^2}{(\sigma_u^2)_0} = \frac{\Delta \sigma_v^2}{(\sigma_v^2)_0} = \frac{\Delta \sigma_w^2}{(\sigma_w^2)_0} \approx \frac{4h}{L} \quad (5)$$

where $\Delta \sigma_i$ is the change from the upwind value $(\sigma_i^2)_0$, $i=u,v,w$. In the outer region the turbulence scales are larger and the eddies are being distorted by the changes in mean velocity. The changes in turbulence may be estimated from rapid distortion theory, yielding

$$\frac{-\Delta \sigma_u^2}{(\sigma_u^2)_0} \sim \frac{\Delta \sigma_w^2}{(\sigma_w^2)_0} \sim \frac{\Delta U}{U_0} \quad (6)$$

Several field experiments have been designed to observe the topographic influence upon surface layer air flow, such as e.g. those on Brent Knoll (Mason and Sykes, 1979), Black Mountain (Bradley, 1980), Ailsa Craig (Jenkins et al., 1981), Nyland Hill (Mason, 1986), Askervein (Taylor and Teunissen, 1988) and Great Dun Fell (Gallagher et al., 1988). These are all relatively large hills, the heights being in the range 70-847 m. Observations over minor topography have been reported by Petersen et al. (1980) and by Smedman and Bergström (1984).

3. EXPERIMENTAL DETAILS

The present site, Ringhals, is located about 60 km south of Göteborg on the west coast of Sweden, and consists mainly of bare rocks with only sparse vegetation (lichen, grass, a few bushes and trees) and some meadow ground in the valleys. The rocky hills are rather inhomogeneous and the topography is shown in Fig. 2. The largest elevation is just about 25 m a.s.l. and the instrumental tower (pos. 1) is located on a smaller hill, 12 m a.s.l., about 1 km from the outermost shore-line. Cup-anemometers, and

ventilated and radiation shielded resistance thermometers were mounted at heights of 7, 15, 21, 31, 43 and 61 m above the surface.

The turbulence data were obtained using a wind-vane based hot-film instrument, which measures the three components of the wind and the temperature at a sampling rate of 20 Hz. Measuring heights were 8, 30 and 63 m above the surface. At 15 m a wind-vane based propeller anemometer of the Gill-type were used.

To get information on the wind and turbulence at other locations, a kite-borne wind-sonde system was used. The wind-sonde is a cup-anemometer whose output is transmitted to a ground based receiver. The sonde data are evaluated to give mean values and standard deviations of the wind speed. Comparisons with instruments mounted on a tower have shown that the mean values are accurate to within a few percent. The standard deviations was corrected 12% as the result of a comparison made at 11 m height between the wind-sonde and the hot-film instrument.

Also the data from the Gill-anemometer was corrected before they were used. Comparing σ_u -values measured with the hot-film instrument and with the Gill-anemometer, the frequency loss was found to be about 15%.

The data analysed here are from the period 2 to 9 October 1987, with intense measurements during the 7th and the 9th when the kite-system was used. During those days, the general weather situation was such that low-pressure systems approached the area from south-west. The weather was sunny in the mornings. During the days, first Ci and Cs, later As and Ns, moved from south-west in over Sweden, followed by rain in the evenings. The wind was turning from about 230° in the morning to 180° in the afternoon.

As seen from Figure 2, the fetch at the tower will differ quite a lot with winds turning from 230° to 180°, and the data have been divided into the three sectors, shown on the map. Surface profiles representing these three sectors are shown in Figure 3. In Sector I the air flow first encounters a 25 m high hill, then passes a valley before coming up the hill to the tower. In Sector II there is only a minor disturbance of the outermost part of a peninsula south of the site. We will regard this sector as 'ideal', and assume that an undisturbed marine boundary layer enters the tower-hill. Finally in Sector III, the air flow approaching the tower is already modified by the upstream passage of the southern peninsula. Also the neighbouring hill is somewhat complex in this sector, having a peak and a minor valley upstream from the tower.

4. RESULTS AND DISCUSSION

We will begin by looking at Sector II, the most 'ideal' sector. The distance from the shore-line to the tower is here about 250 m, giving a length-scale $L \approx 125$ m. The height of the hill is 12 m and the roughness length about 0.1 m, giving an inner layer height $l \approx 7$ m, and $\Delta S_{max} \approx 0.15$.

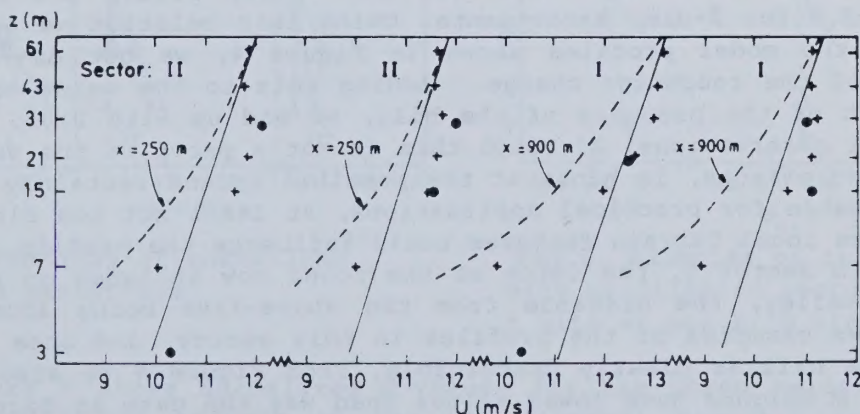


Fig.4. Examples of wind profiles; + tower; • kite at pos.2; — model over the sea; --- model at given distances from the shore-line.

Examples of wind profiles in this sector are shown in Figure 4. Above 15 m the gradient is very small, which is an indication of the influence from the hill. The temperature profile was very close to neutral during the whole experiment. The kite data (marked •, taken at pos.2, Fig.2) agree reasonably well with the tower data. One must, however, be careful about how well these data represent the conditions over the sea, as the hilly coast may influence the wind field already at the shore-line. The data at 3 m was in fact not taken with the sonde kite-borne, but instead it was mounted on a small mast just at the water's edge. The string of the other sonde(s) was fastened close to this mast, but as the height angle of the string was about 40°, this meant that this sonde was about 30-40 m inland, where it could definitely be influenced by the hill. Model results from Taylor and Lee (1984, Fig.7-8) suggest that the data from 3 m may be ca. 5-10% too high for these wind directions.

In Figure 4, logarithmic wind profiles representative of the sea are also included, and we see that the 7 m level is affected by the change in surface roughness. The model used is described in Bergström (1986), and consists of a surface layer model based upon the Monin-Obukhov similarity theory, and above this an Ekman-Taylor type model with an eddy-diffusivity constant with height. Tests against field data have shown this procedure to be adequate when modelling mean profiles. The model also includes a parameterized scheme to calculate the growth of internal boundary layers, following a surface discontinuity. In Figure 4 the modelled profile at the distance 250 m from the coast-line is included, and we see that the 7 m wind speed is in between the upstream value and the value predicted by the model at this distance from the shore-line, indicating a combined effect of speed-up due to the hill and retardation due to increased roughness.

Now let us try to quantify this. In Figure 5, wind profiles normalized with the value at 61 m are shown, together with a logarithmic profile representing the sea ($z_0=0.00025$ m). At 60 m the two gradients agree very well, while in the layer 10 to 40 m the 'speed-up' caused by the hill is clearly seen. At 7 m, which is about the height of the inner layer, the change in wind speed seems to be dominated by the change in surface roughness. A question which then arises is whether these two effects could be added. To test this we use the observations together with model results at 15 m above the surface. From Figure 5 we get the fractional speed-up ratio 0.05 at this height, and from Eq.(4) we get $\Delta S_{\max}=0.15$. Now the 15 m level is well above the height of the inner layer, thus we should expect to get $\Delta S < \Delta S_{\max}$. Taylor and Lee (1984) has suggested

$$\Delta S(z) = \Delta S(0)e^{-Az/L} \quad (7)$$

as an approximate relation, where $A=4$ for 3-dim. hills, $A=3$ for 2-dim. hills and $A=2.5$ for 2-dim. escarpments. Using this relation we get $\Delta S(15)=0.09$. Using the model profiles shown in Figure 4, we get $\Delta S=-0.04$ as a consequence of the roughness change. Adding this to the value expected to be the result of the presence of the hill, we end up with 0.05, in agreement with the observations. Although this is not a proof of the validity of adding the two effects, it hints at the possibility and certainly makes the procedure usable for practical applications, at least not too close to the surface, where local terrain features could influence the profile.

Turning to Sector I, the fetch at the tower now includes an additional hill and a valley, the distance from the shore-line being about 900 m. Figure 4 shows examples of the profiles in this sector, and once again the effect of the hill is clearly discernable. From Figure 5 we also see that the 7 and 15 m heights have lower values than was the case in Sector II, in accordance with the longer travel distance over land. Estimating the fractional speed-up ratio at 15 m as above, ignoring the upstream hill, we end up with $\Delta S=0.00$, while the observations give $\Delta S=0.02$, in approximate agree-

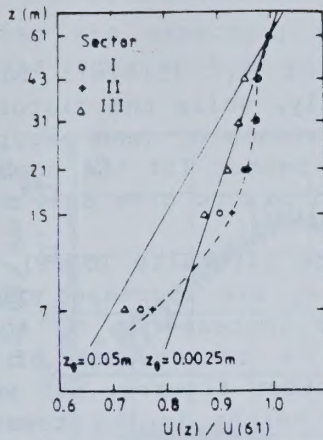


Fig.5. Observed normalized mean wind profiles and together with modelled profiles at two roughness lengths; — model; --- obs.sector II.

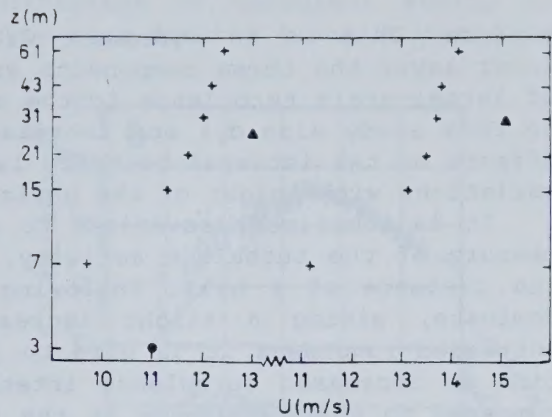


Fig.6. Examples of wind profiles in sector II; + tower; • kite at pos.2; Δ kite at pos.3.

ment with the results of Counihan et al. (1974). They made wind tunnel studies over 2-dim. ridges, and found a significant speed up ($\Delta S \approx 0.4$) on the first hill, while on the second and subsequent hill $\Delta S < 0.1$.

Finally the wind profiles from Sector III, included in Figure 5, show a mixture of flow-over-hill and internal boundary layer effects. The fact that the mean profile in this sector does not deviate much from the marine wind profile is probably a combined effect of the retardation over the southern peninsula and the subsequent acceleration over the cove and speed-up due to the nearby hills.

On a hill closer to the coast-line (pos.3 in Fig.2) the wind-sonde was used to get some additional information. These data are plotted in Figure 6, together with the corresponding tower data. As may be seen from the figure, at the location of the sonde the wind speed is much higher than at the tower, while when the sonde was used at the shore-line the two agreed quite well. Estimates of the fractional speed-up ratios, following Eqs (4) and (7) gives 0.09-0.10, which is in agreement with the observations.

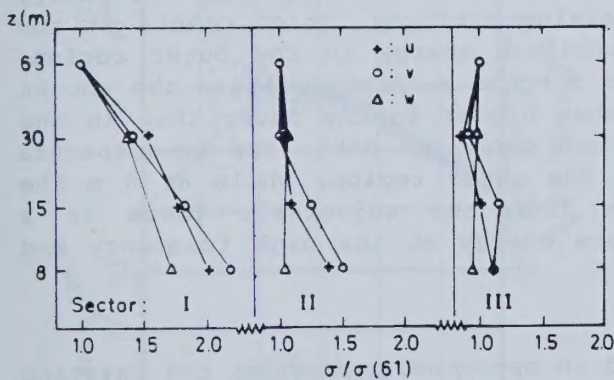


Fig.7. Profiles of normalized standard deviations.

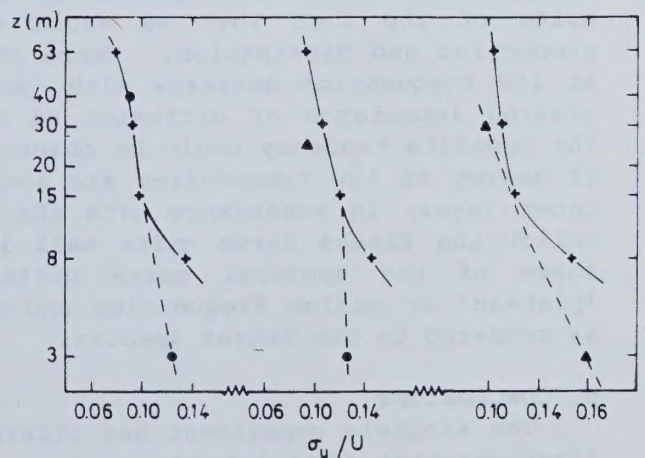


Fig.8. Examples of turbulence intensity profiles in sector II; + tower; • kite at pos.2; Δ kite at pos.3.

Figure 7 shows normalized σ -values for the three velocity components. In Sector I the effects of the new internal boundary layer dominates, giving an increase in all three components. In Sector II σ_w is practically constant with height, whereas both σ_u and σ_v increase closer to the

surface. This is in agreement with Hunt (1980), cf Eq. (5)-(6). In the inner layer the three components are affected equally, while the distortion of larger-scale turbulence in the outer layer will reduce σ_u (and according to this study also σ_v) and increase σ_w . Finally in Sector III the combined effects of two internal boundary layers and the hill mix to give only minor variations with height of the horizontal components.

It is sometimes convenient to use the turbulence intensity (σ_u/U) as a measure of the turbulent activity. Now both U and σ_u are increased due to the presence of a hill. Following Eqs (4)-(5) the increase in σ_u should dominate, giving a slight increase in σ_u/U . As a consequence of the increased roughness it is hard to judge about the exact figures, but we do find an increased turbulence intensity at the 8 m height in the tower as compared to the wind-sonde at the shore-line, see Figure 8. The wind-sonde data from the hill at position 3 in Figure 2 are also included, and they agree fairly well with the tower data.

Following Panofsky et al. (1982), assuming balance between local production and dissipation, the velocity spectra, $S(n)$, in the inner layer could, be normalized according to

$$nS(n) / (-\overline{u'w'} \cdot u_{*p})^{2/3} = 0.30f^{-2/3} \quad (8)$$

where $\overline{u'w'}$ is the local Reynolds stress and u_{*p} is the friction velocity, evaluated from the wind profile assuming this to be logarithmic. The dimensionless frequency $f=nz/U$. Over uniform terrain, $-\overline{u'w'}=u_{*p}^2$, and Eq.(8) is the normalization used by Kaimal et al. (1972). Figure 9 shows the u - and w -spectra, together with the uw -cospectra, normalized in this manner. Included are also the corresponding Kansas-curves. The agreement is very good at high frequencies, while at the low frequency end the spectra from Ringhals contain more energy, in contradiction to what Panofsky et al. (1982) found when analysing data from Black Mountain. The reason for this is likely to be found in differences in the filtering techniques, as only a mean value filter was used when calculating the spectra from Ringhals. The same result was found at Maglarp (cf. Smedman and Bergström, 1984). It is also surprising to note that this normalization seems to work in the outer region as well. This was also found by Panofsky et al. (1982) for the w -spectra, but here the u -spectra as well follow the inner layer scaling, in spite of the fact that we would not expect equilibrium between local production and dissipation. There is, however, a tendency that the energy at low frequencies decrease with increasing altitude, which points at the growing importance of diffusion of turbulent energy in the outer region. The opposite tendency could be observed for the w -spectra, where the amount of energy at low frequencies are somewhat higher in the outer than in the inner layer, in accordance with the Black Mountain data. The uw -cospectra follow the Kansas curve quite well in the outer region, while at 8 m the shape of the spectral curve differs from the universal. There is a 'plateau' at medium frequencies and more energy at the high frequency end as compared to the Kansas spectra.

5. CONCLUSIONS

The Ringhals experiment has offered an opportunity to test the existing flow-over-hill theories at a site with low, rather inhomogeneous rocky hills with sparse vegetation. It has been found that when the internal boundary layer is taken into account, the data agree as well as could be expected with the theoretical predictions. Thus, in spite of the increased roughness, the wind speed at 10-30 m height above the surface is often found to be higher over the 10-20 m high hills than over the adjacent sea.

Also the turbulence seems to follow the flow-over-hill theories, both as regards the σ -profiles and the shape of the spectra in the inner layer. In the outer region spectra agree better with those from 'ideal' sites

than was expected, suggesting that diffusion of turbulent energy is of less importance at this site than was the case for the Black Mountain data.

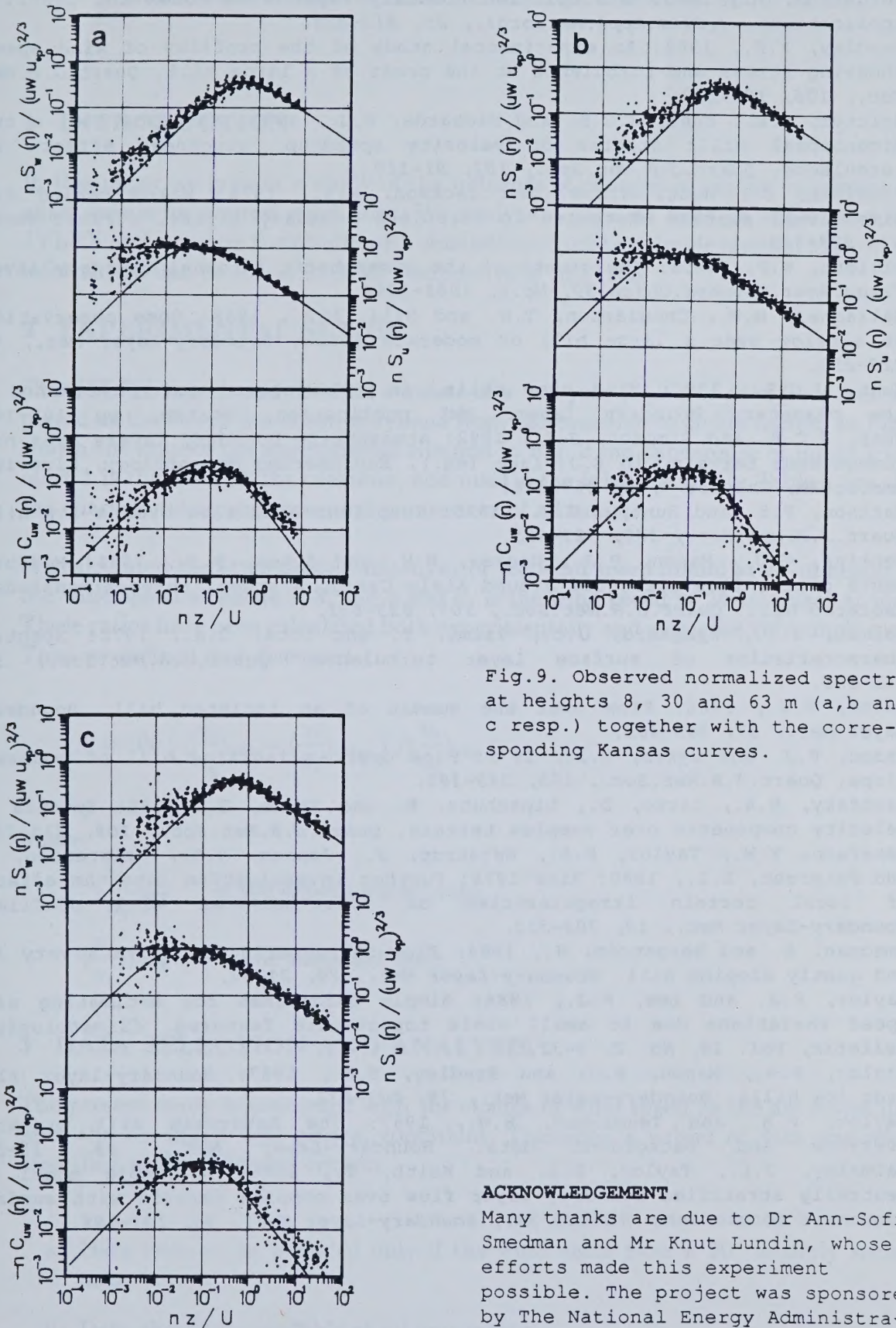


Fig.9. Observed normalized spectra at heights 8, 30 and 63 m (a,b and c resp.) together with the corresponding Kansas curves.

ACKNOWLEDGEMENT.

Many thanks are due to Dr Ann-Sofi Smedman and Mr Knut Lundin, whose efforts made this experiment possible. The project was sponsored by The National Energy Administration in Sweden.

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ROUGHNESS CHANGE EFFECTS FOR SMALL AND LARGE FETCHES

Anna Maria Sempreviva, S.E. Larsen, N.G. Mortensen and I. Troen

1 INTRODUCTION

In this paper we present a study of the influence of changes in surface roughness z_0 on wind speeds for neutral conditions. We analyze a data set from an experiment called JYLEX and compare the results with simulations from two simple models. More detailed information can be found in A.M. Sempreviva et al., 1988.

2 EXPERIMENTAL SET-UP

The JYLEX experiment data set consists of a two-year period of data from four masts placed at the North Sea coast of Jylland from the coastline to 30 km inland. In Fig. 1 is shown the map of the experimental site and in Fig. 2 the appearance of masts 1 and 4; mast 1 is the mast at the coastline, and mast 4 the inland mast. In Table 1 are shown the characteristics and measured parameters of all masts.

We calculated the attenuation ratio u_i/u_1 of the wind speed inland at 24-m height. u_i is the wind speed at masts 2, 3, and 4 and u_1 is the wind speed at 24-m height at mast 1. These ratios have been calculated both experimentally and using the two simple models. Then we defined and discussed

$$\delta = 100 \left(\left\langle \frac{u_i}{u_1} \right\rangle_{exp} - \left\langle \frac{u_i}{u_1} \right\rangle_{comp} \right) / \left\langle \frac{u_i}{u_1} \right\rangle_{exp} \quad (1)$$

where

$\left\langle \frac{u_i}{u_1} \right\rangle_{exp}$ is the experimental ratio

$\left\langle \frac{u_i}{u_1} \right\rangle_{comp}$ is the modelled ratio

3 DATA SELECTION AND ANALYSIS

The present study is concerned with the change of wind speed as the air moves inland from the sea under near-neutral conditions. Therefore, a subset of data was selected according to the following criteria:

- a) Data were to be included only if the wind came from a 90° westerly sector at mast 1.
- b) Data should be available at all four masts.

Table 1: The JYLEX experiment. For each of the four masts are shown distance to the coast, height, and number of measurement levels for the various parameters.

Station	Mast 1	Mast 2	Mast 3	Mast 4
Distance to coast [km]	0.08	1.2	4.4	30.2
Height of mast [m]	32	24	24	24
Wind speed	6	3	3	6
Wind direction	2	2	2	2
Gust wind speed	1	1	1	1
Temperature	2	2	2	2
Temperature gradient	4	2	2	4
Relative humidity	3	1	1	3
Precipitation			1	
Atmospheric pressure	1			1
Incoming short-wave	1		1	
Sonic anemometer			1	

- c) Near-neutral data were ensured by demanding a wind speed larger than 12 m/s at the top level of mast 1 while at the same time the absolute value of the Richardson number should be less than 0.03 at all masts. The Richardson number used refers to 10-m height.

The data set selected in this way consisted of 2048 sets of profile data recorded simultaneously along each mast, meaning that 2 per cent of the data fulfilled the above criteria. It was stratified subsequently according to the following criteria.

1. Day or night
2. Season: winter (December, January, and February
spring (March, April, and May)
summer (June, July, and August)
fall (September, October, and November).
3. Finally, the 90° direction sector was subdivided into nine 10° sectors.

The most important reason for introducing the seasonal criterion is that the roughness over land varies with season following the vegetation and other aspects of the surface such as snow cover and tilling of the farmland.

The subdivision into 10° direction sectors was made because it allowed us to determine fairly well-defined fetch conditions for each mast. The appropriate wind direction sectors were determined on the basis of the measurements at mast 1. However, during the strong wind conditions considered here it was found that all masts always showed a

wind direction within the same 10° sector. Figure 3 illustrates the direction sectors for mast 3.

Between the velocity u_i at mast i ($i = 2, 3, 4$) and the upstream over-water velocity u_1 the ratios were calculated for each record at the 24-m level. Subsequently, the average values and standard deviations of these ratios were computed within each bin defined by the season and wind direction criteria given above.

As said above, the upstream wind was determined from mast 1. Due to the presence of an approximately 100-m wide rush field in front of mast 1, we used the 31-m wind to estimate the over-water wind at the height of 24 m. This was done using Charnock's relation in conjunction with a logarithmic wind profile.

$$\begin{aligned} u_* &= \kappa u_{31} / \ln(31/z_{0w}) \\ z_{0w} &= c u_*^2 / g \\ u_{24} &= \frac{u_*}{\kappa} \ln(24/z_{0w}) \quad \text{with } c = (1.4 \times 10^{-2}) \end{aligned} \tag{2}$$

Having determined the fetch to the water, the sector and seasonal averages of u_i/u_1 , the corresponding standard deviations can be plotted versus land fetch. This is done in Fig. 4 for the winter and summer data.

The velocity ratio is generally seen to decrease with increasing fetches. However, there is considerable scatter. This reflects that plotting $\langle u_i/u_1 \rangle$ versus fetch only is a strong idealization. In reality, the velocity at each mast reflects the upstream history of the flow, and with few exceptions a trajectory passing one mast will not pass any of the others. However, in the following we shall use the idealized picture at least to the extent that we will use the velocity at mast 1 as upstream over-water conditions for all masts.

Figure 4 shows also that the standard deviation of each $\langle u_i/u_1 \rangle$ increases with increasing fetch. This can simply be explained by noting that the correlation between u_1 and u_i fluctuations is getting smaller for the larger distances involved in spite of the 10-min averaging employed.

As already indicated in Fig. 4, winter data are generally high and summer data low while the spring and fall data are less clear. As mentioned above, this behaviour probably reflects a mostly vegetation-controlled variation of the land roughness. As a detail, we note that the winter value is below average for mast 2. This undoubtedly reflects the cycle of growing and harvesting of the 100-m rush zone in front of mast 1. This zone is harvested at the end of February and in the beginning of March. The rush grows to a height of about 2 m in late summer and remains at this height during winter. It is likely that this rush zone does not influence the 31-m velocity at mast 1 (at least not for sectors 2–8), but it certainly influences the velocity at mast 2, being 1 km downwind from this zone.

4 MODEL DESCRIPTION

For comparison with the data, we present here two simple models for describing the flow response to step changes in surface roughness; one pertaining to the surface layer only (SL) while the other includes the boundary layer height as well (BL).

The first part of the model is described in Hedegaard and Larsen (1982) and Larsen et al. (1982); originally, this model comes from Miyake (1965) and is further discussed in Panofsky (1973), Businger (1972) and Jensen (1978)). The basic equation of the BL model is the following.

When the flow passes a change in surface roughness, an internal boundary layer grows as

$$\frac{\partial h}{\partial x} = A \cdot \frac{\sigma_w(h)}{u(h)} \quad (3)$$

in which h is the height of the internal boundary layer, x is the fetch downwind of the roughness change, while u is the mean speed and σ_w the standard deviation of the vertical wind speed. The two last parameters are described by

$$\begin{aligned} u &= \frac{u_{*0}}{\kappa} \left(\ln \frac{z}{z_0} - 2 \frac{z}{H} \right) \\ \sigma_w / u_{*0} &= \frac{\sigma_w}{u_{*0}} \left(1 - \frac{z}{H} \right)^2 \\ H &= u_{*0} / f \end{aligned} \quad (4)$$

in which f is the Coriolis parameter, κ the von Kármán constant, z_0 the roughness length, and H the scale height. Subscript 0 indicates that the parameter refers to the surface.

The SL theory is recovered for $\frac{z}{H} \ll 1$.

In Fig. 5 is shown the growth of an IBL in a two-dimensional PBL for smooth to rough transition. In Fig. 6 is shown the profile used in the IBL. We divide the IBL in three zones delimited from three characteristic heights (Jensen and Peterson (1978)): $h(x)$ is the IBL height, $h_1(x) \sim \frac{1}{3}h(x)$ and $h_2(x) \approx \frac{1}{15}h(x)$. Let h_c be the height where we calculate the downstream wind speed.

$$\text{If } h_1(x) < h_c < h(x)$$

the profile parameters belong to the upstream conditions;

$$\text{if } h_2(x) < h_c < h_1(x)$$

the wind speed is calculated by a linear interpolation of the upstream and downstream wind profiles;

$$\text{if } h_c < h_2(x)$$

the profile parameters belong to the downstream condition.

4 DATA INPUT FOR THE MODELS

- wind speed at 31-m height at mast 1 for each direction,
- fetches to the water or the the next change of z_0 for each mast and direction,
- Charnock's constant for the calculation of z_0 over sea,
- z_0 over land corresponding to each fetch (Fig. 3),
- for long fetches z_0 was estimated from topographic maps; around the mast z_0 was calculated from the measured wind profile.

5 COMPARISON BETWEEN MODEL AND DATA

For a detailed comparison we computed δ in Eq. (1) for each mast 2, 3, and 4, sectors 2 – 8, and season.

As seen before, the choice of only sectors 2 to 8 is related to the narrow belt of rush which grows until 2 m along the coastline. The roughness of this rush was determined by the data from sector 9 which was the only sector, maybe together with sector 1, with sufficient rush fetch in the front.

Thus to be sure that the 31-m level of mast 1 was not involved in the rush IBL we dropped sectors 1 and 9.

As discussed in the preceding section, the estimate of the surroundings of the roughness at each mast is associated with quite some uncertainty. Therefore, we cannot test in a strict sense the absolute validity of the model approaches considered. Instead we will address the following uncertainties:

- a) How will δ change with a changing estimate of the water roughness?
- b) What is the influence on δ when using the extended BL model rather than the surface layer model?
- c) How does introduction of the kinked profile in Fig. 6 influence δ ?

As a basic model we choose the extended BL model given by Eq. (6). The kinked profile is used with $h_1 = h/3$ and $h_2 = h/15$. The upstream water roughness is determined by Charnock's constant $c \sim 1.4 \times 10^{-2}$. Figures 7, 8 and 9 illustrate how δ changes when

we change one of the aspects considered under questions a) to c) in turn, keeping the other model characteristics as in the basic model.

First, we study the influence of changing Charnock's constant. The results are summarized in Fig. 7. The c -value producing $\bar{\delta}$ close to zero for all three masts is seen to be between 1.4×10^{-2} and 4.2×10^{-2} around $c \sim 3 \times 10^{-2}$. This value is somewhat larger than the "normal" value 1.4×10^{-2} . However, the nearest part of the upstream conditions is either the shallow fjord or the coastal water (see Fig. 1). It is therefore not surprising to find z_0 somewhat larger than the "open-ocean" value (see e.g. Geernaert et al., 1987).

We can relate the variation of δ to both changes in z_0 and to Charnock's constant c -variation. Note that relative changes in water roughness equal the relative changes in c .

For simplification we shall use the surface layer model only, neglecting the profile kinks. Also, we limit the description to two roughnesses only, z_{01} pertaining to water, and z_{02} describing the land roughness.

The modelled ratio between upstream and downstream velocity is found from Larsen et al. (1982)

$$\frac{u_2(z)}{u_1(z)} = \frac{\ln \frac{h}{z_{01}}}{\ln \frac{h}{z_{02}}} \cdot \frac{\ln \frac{z}{z_{02}}}{\ln \frac{z}{z_{01}}} \quad (5)$$

Based on the Eqs. (1) and (5) we can find the changes in δ related to changes in z_{01} and z_{02}

$$d\delta = a \frac{dz_{02}}{z_{02}} - b \frac{dz_{01}}{z_{01}} \quad (6)$$

with

$$a = \frac{100}{R_s} \frac{u_2}{u_1} \frac{\ln \frac{h}{z}}{\ln \frac{h}{z_{02}} \ln \frac{z}{z_{02}}} \quad (7)$$

$$b = \frac{100}{R_s} \frac{u_2}{u_1} \frac{\ln \frac{h}{z}}{\ln \frac{h}{z_{01}} \ln \frac{z}{z_{01}}}$$

where R_s is the experimentally determined ratio between u_2 and u_1 , compare Eq. (1).

With average land fetches of the order of 45, 7 and 2 km for masts 4, 3, and 2, respectively, and with $z_{02} \simeq 20$ cm and $z_{01} \simeq 0.1$ mm, the estimates of the a 's and b 's are summarized in Table 6 for the different masts.

In Fig. 7 we present the influence of the estimated upstream water roughness by plotting the yearly average values of $\bar{\delta}$ at the three masts for different values of Charnock's

Table 2: Estimates of the sensitivity factors a and b in Eq. (6) for mast 2, 3, and 4.

Mast	a	b
2	5.3	1.02
3	7.6	1.5
4	9.8	2.2

constant, c . Note still that the relative change in water roughness equals the relative change in c .

The influence of Charnock's constant c is shown, using both the extended BL and the SL-model. From the figure is seen that for all three masts the data model comparison is internally consistent in showing that $\bar{\delta} \sim 0$ for $c = 3.0 \times 10^{-2}$ as previously noted. For the SL-model the δ -values at mast 4 look slightly less consistent with those for the two other masts than for the BL-model. However, from Eq. (6) and Table 2 is seen that a 15% reduction of large-scale roughness, to which mast 4 is more sensitive than the two other masts, would reverse the picture.

The figure also illustrates that the sensitivity of $\bar{\delta}$ to changes in Charnock's constant increases with distance to the shore line as predicted by Eq. (6) and Table 2.

Next we shall study the influence on δ of changing the height, h_1 , down to which the outer profile is supposed to describe the resulting profile. Here, our basic situation is $h_1 \sim h/3$, and we shall see the effect of increasing this height to $h/2$ and h , where the last value corresponds to the velocity profile being found just by matching the inner and outer profile in $z = h$.

In Fig. 8 is illustrated the influence of $\bar{\delta}$ of using different values of h_1/h . The difference between the δ 's at the three masts increases for $h_1/h > 0.5$. Equation (6) and Table 2 show that this effect cannot be repaired by changing Charnock's constant. Hence, we conclude that the kinked profile with $h_1/h = 0.3$ does indeed improve the performance of the models. This latter point is illustrated in Fig. 9 which corresponds to Fig. 8, but with Charnock's constant $c = 3.0 \times 10^{-2}$.

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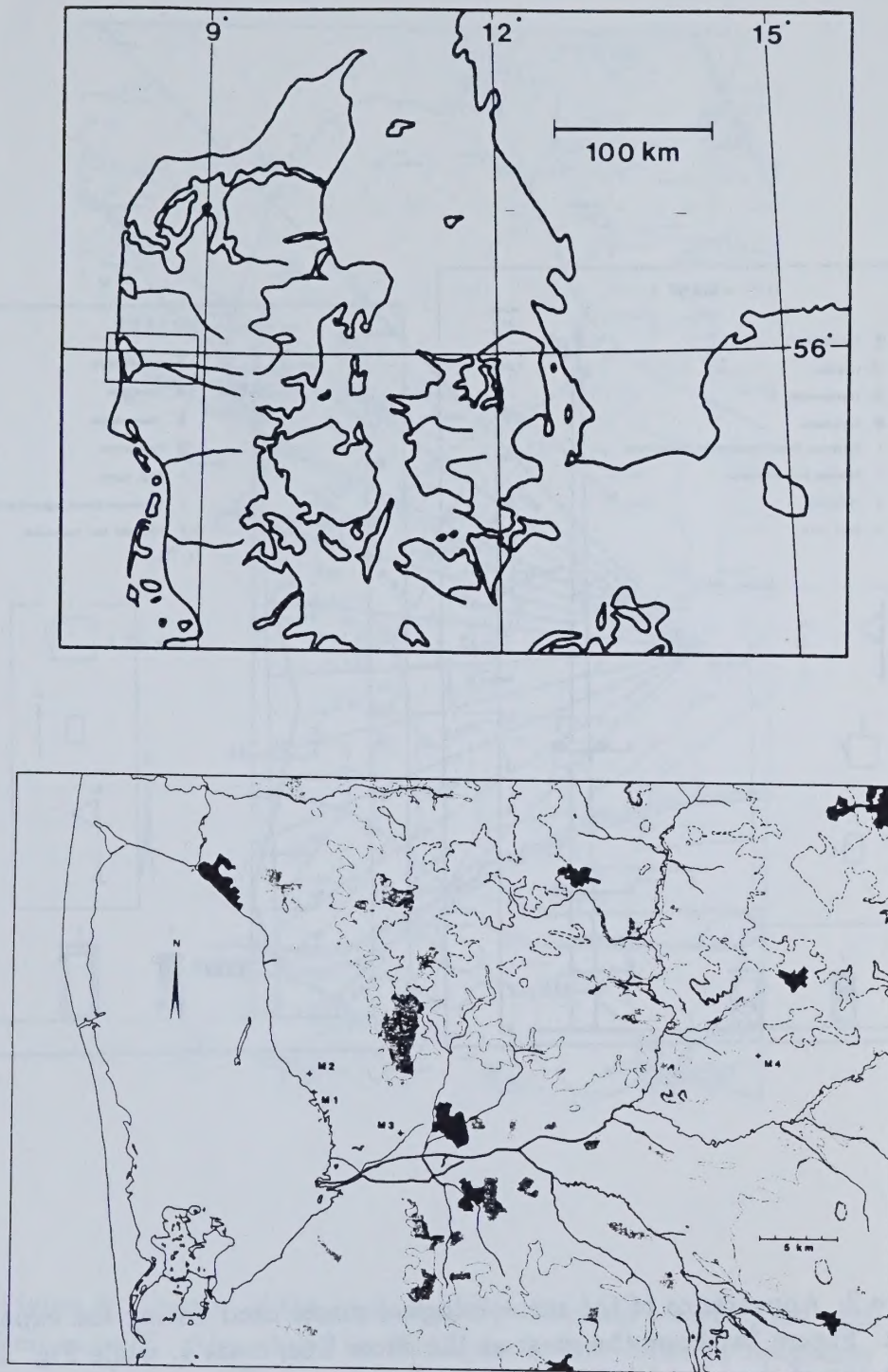


Figure 1: Maps of the experimental site. Figure 1a shows the overall area, while Fig. 1b gives a more detailed map of the site, indicating positions of the masts. In Fig. 1b main terrain features are also indicated such as cities, forests, and heights of terrain.

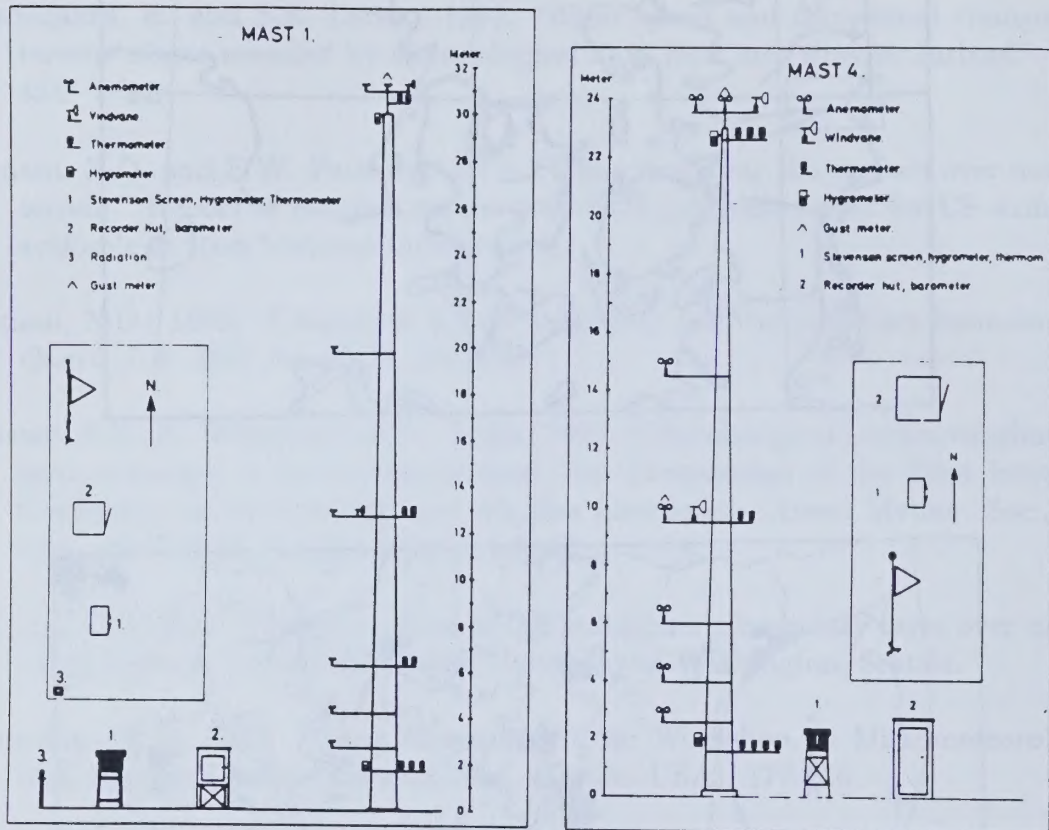


Figure 2: Appearance of the meteorological masts used during the experiment. Figure 2a shows the mast at the shore line, mast 1, while Fig. 2b shows one of the inland masts, mast 4.

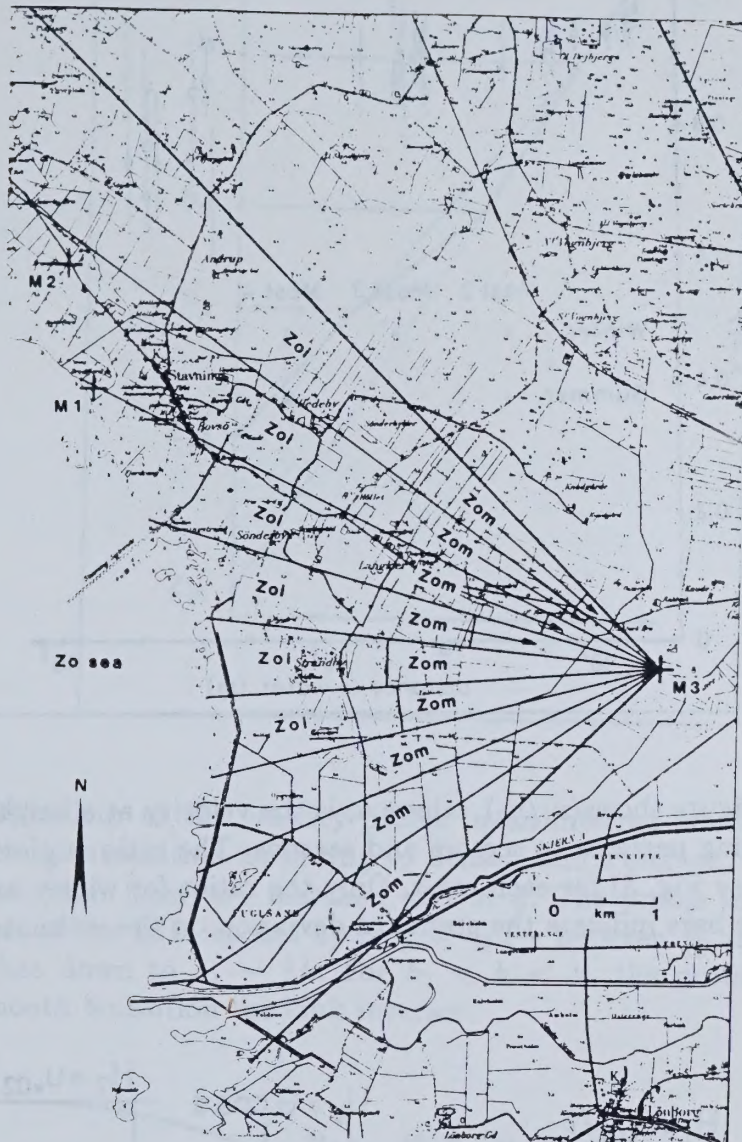


Figure 3: Layout of the nine westerly direction sectors from each mast, here mast 3. Also shown in the figure is how different roughness values are ascribed to different areas for use in the model computations. z_{0m} -values are estimated from the profile measurements, while the z_{0i} -values are estimated as described in the text. Many of the z_{0i} -areas in the figure are further subdivided into areas with different z_{0i} . For simplicity this is omitted in the figure.

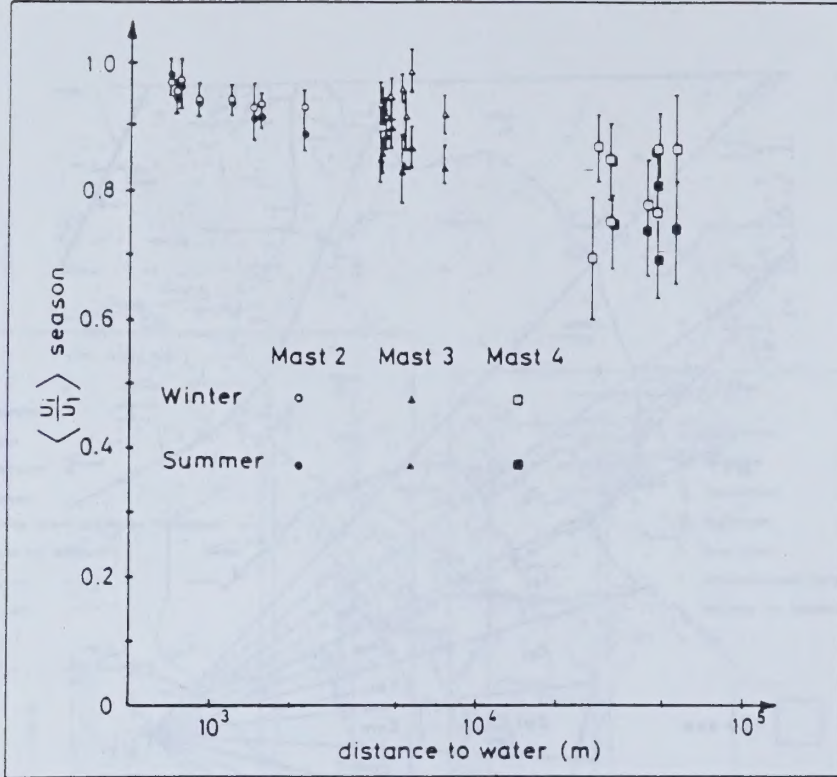


Figure 4: The figure shows $\langle u_i / \bar{u} \rangle$, where u_i is the velocity at a height of 24 m at mast i and the averaging pertains to sectors and seasons. The ratio is plotted versus distance to the water (see Fig. 3) for each mast. Only the ratios for winter and summer seasons are shown. The bars indicate the standard deviation on the estimated $\langle u_i / \bar{u} \rangle$.

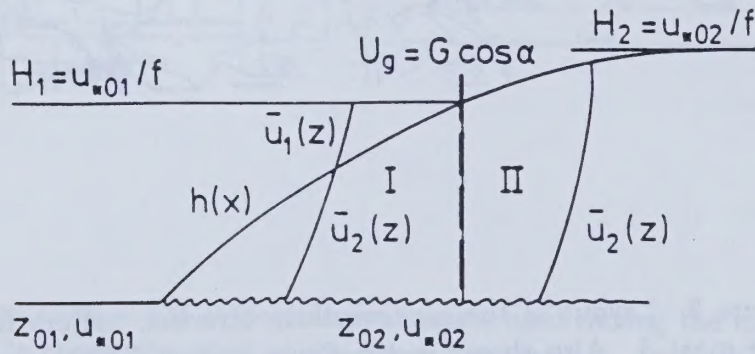


Figure 5: Growth of an internal boundary layer (IBNL) in a two-dimensional planetary boundary layer (PBL) for smooth-to-rough transition. In Zone I the IBL grows within the smooth PBL, while in Zone II $h(x)$ is above the smooth PBL (Larsen et al., 1982).

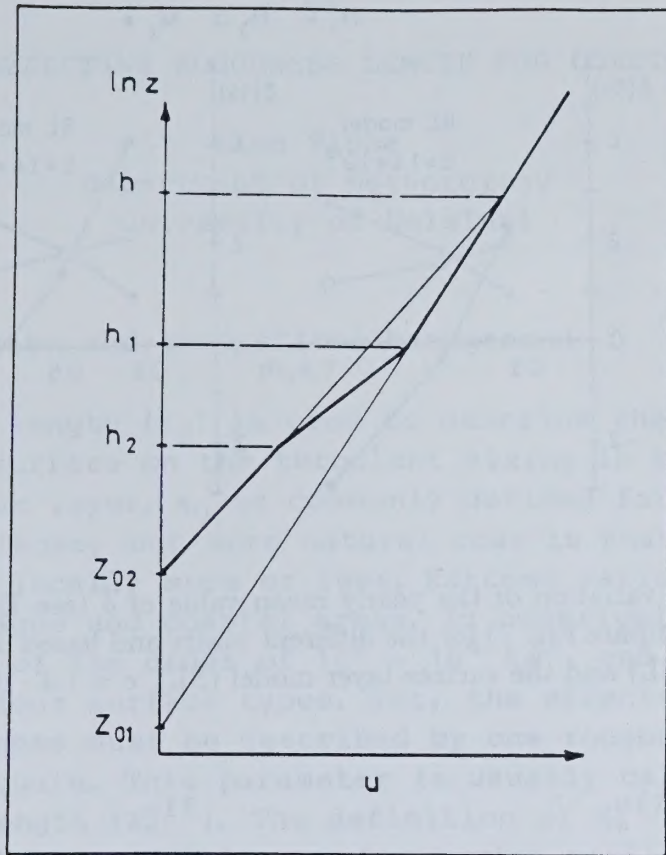


Figure 6: Behaviour of the velocity profile in an internal boundary layer (here for a smooth-to-rough transition according to Jensen and Peterson, 1977). The outer and inner profiles (thin lines) are matched at $z = h$. The profile in equilibrium with the surface stress of the IBL reaches up to $z = h_2$ ($\sim$ between $\frac{1}{10}$ and $\frac{1}{20}$ of h). The outer profile reaches down to $h_1 \sim \frac{1}{3}h$. For $h_2 < h < h_1$ the profile is interpolated. For rough-to-smooth transition the kink reverses.

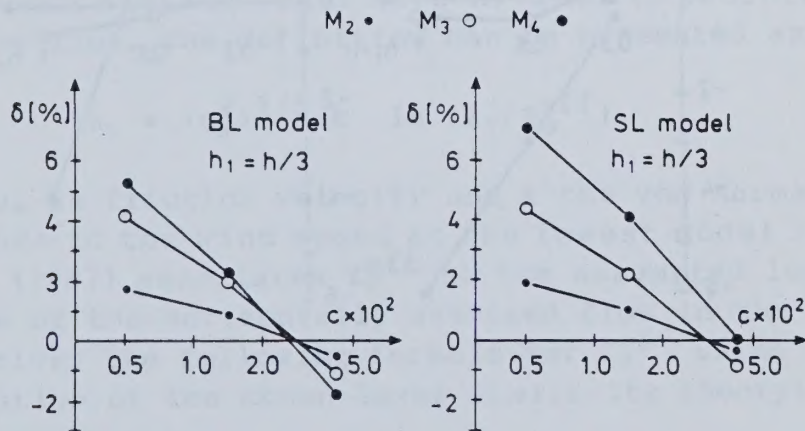


Figure 7: The variation with Charnock's constant, c , of the yearly mean values of δ (see Eq. (1)) for the different masts based on both the full boundary layer model (IBL) and the surface layer model denoted SL.

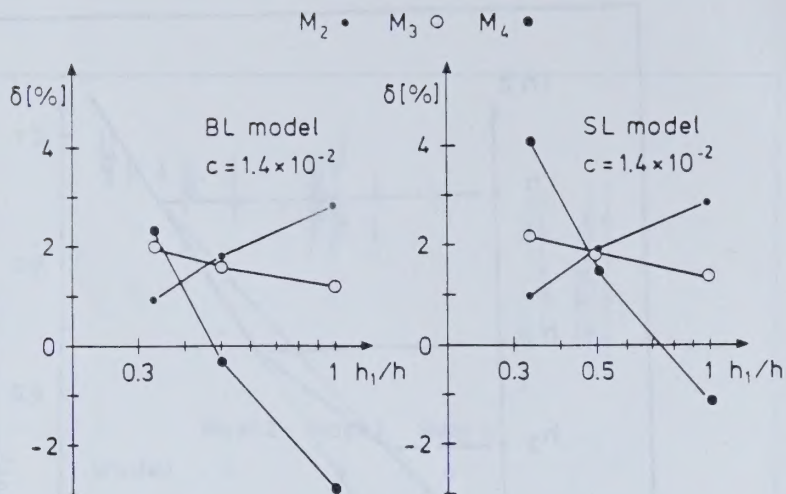


Figure 8: The variation of the yearly mean value of $\bar{\delta}$ (see Eq. (1)) versus the height ratio h_1/h (compare Fig. 7) for the different masts and based on both the full boundary layer model (BL) and the surface layer model (SL). $c = 1.4 \cdot 10^{-2}$.

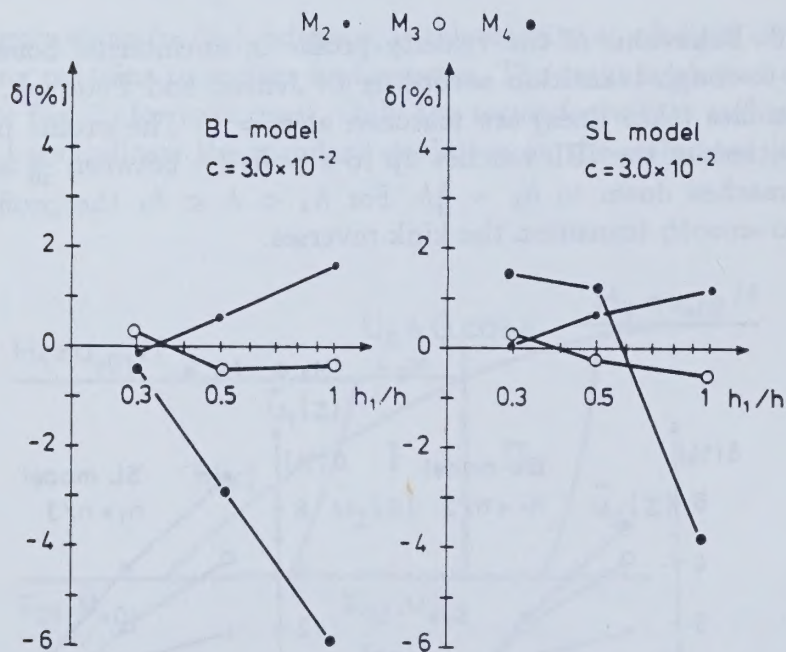


Figure 9: The variation of the yearly mean value of $\bar{\delta}$ (see Eq. (1)) versus the height ratio h_1/h (compare Fig. 8) for the different masts and based on both the full boundary layer model (BL) and the surface layer model (SL). $c = 3.0 \cdot 10^{-2}$.

AN EFFECTIVE ROUGHNESS LENGTH FOR COASTAL AREAS

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1. Introduction and theoretical background

Roughness length (z_0) is used to describe the frictional effects of surface on the turbulent mixing in the atmospheric constant flux layer. z_0 is commonly defined for horizontally uniform surfaces, but more natural case is that surface roughness varies locally more or less. Extreme variations are found in archipelagos and coastal areas. In numerical models grid squares are of the order of $10^2 - 10^4 \text{ km}^2$, and so they usually contain various surface types. Yet, the effects of sub-grid scale roughness must be described by one roughness length in each grid square. This parameter is usually called an effective roughness length (z_0^{eff}). The definition of z_0^{eff} is not simple, as it is not connected to any local wind profile. z_0^{eff} is some kind of average of the local z_0 's in the grid square. Kung (1963) calculated z_0^{eff} 's simply by taking area-weighted average (marked $\langle \rangle$) of $\ln z_0$:

$$\ln z_0^{\text{eff}} = \langle \ln z_0 \rangle \quad (1)$$

Although this is a sound estimate, calculation should have better physical basis if we need more accurate results. Fiedler and Panofsky (1972) defined z_0^{eff} as the roughness length, which homogeneous surface would have in order to produce the correct momentum flux. The definition can be presented as:

$$u_1 = \langle u_*^2 \rangle^{1/2} / k \ln (z_1 / z_0^{\text{eff}}) \quad (2)$$

where u_* is friction velocity and k the von Karman constant. u_1 refers to the wind speed at the lowest model level z_1 . Taylor (1987) associates z_0^{eff} to the assumed logarithmic profile of the horizontally averaged flow in the grid square, and derives the following formula for z_0^{eff} using linear approximation of the Ekman-layer similarity theory:

$$\ln z_0^{\text{eff}} = \langle \ln z_0 \rangle + a \sigma_{\ln z_0}^2 \quad (3)$$

where a is a parameter ($a \approx 0.1$) and $\sigma_{\ln z_0}^2$ is the variance of $\ln z_0$. It has also been suggested (van Dop, 1983; Wieringa, 1986) to calculate z_0^{eff} from the area average of surface layer drag coefficient (e.g. at the height of the lowest model level):

$$(\ln(z_1/z_0^{\text{eff}}))^{-2} = \langle (\ln(z_1/z_0))^{-2} \rangle \quad (4)$$

This scheme would lead to correct momentum flux only if the wind speed is the same everywhere at the calculation level z_1 . Therefore Mason (1986) lets z_1 depend on the fetches in homogeneous surface sections in the grid square ($z_1 = L/400\text{m}$, where L is the fetch).

The above definitions associate z_0^{eff} to the distribution of local micrometeorological z_0 , which stands mainly for surface friction. However, if there are large terrain height variations in the grid square as is often the case in coastal areas, the sub-grid scale form drag should also be parametrized. In such cases z_0^{eff} should be multiplied by a form drag factor, which depends on the spectra of terrain height variations (c.f. Mason, 1986; Emeis, 1987).

2. Model Simulations

A two-dimensional hydrostatic mesoscale boundary-layer model (Alestalo and Savijärvi, 1985; Savijärvi and Alestalo, 1988) was used to study different theories of calculating z_0^{eff} . The whole model area (265 km, 65 grid points) simulated a grid square of a large scale model and the mesoscale model produced the sub-grid wind field for a given z_0 -distribution using 10 m/s geostrophic wind and neutral lapse rate. The model was used diagnostically to let the balanced wind field settle for a given z_0 -field, and the aims were to find out the best functional dependence between z_0^{eff} and z_0 -distribution, and to investigate the effects of internal boundary-layers and the height of the lowest model level. Two values of local z_0 were used in the model: 0.001 m for water and 0.5 m for land surface.

The area was usually divided into three islands, whose size and so the amount of water surface varied in different simulations. z_0^{eff} was calculated from the profile of the horizontally averaged flow, and its dependence on the percentage of land surface is given in figure 1. Values calculated from the theories are also shown. It seems obvious that the inaccuracy of determining z_0^{eff} is greatest when z_0^{eff} differs much from both land and sea values. It can also be seen that theories, which give the correct momentum flux or presume logarithmic mean flow profile, correspond rather well to the model results. In addition, simulations were done with various numbers of islands to study the effects of internal boundary layers. z_0^{eff} was found to increase when the fetches in the homogeneous surfaces de-

crease. The square root of the wave number of surface variations (k , $k = 2\pi/\lambda$, $\lambda = 2 \cdot$ homogeneous fetch) seemed to be proportional to $\ln z_0^{\text{eff}}$, and a following parametrization was developed for coastal areas:

$$\ln z_0^{\text{eff}} = \langle \ln z_0 \rangle + a \sigma_{\ln z_0}^2 + 1.7 k^{1/2} \quad (5)$$

where k is expressed in km^{-1} . (5) has been verified only by model simulations in areas where $z_0^{\text{eff}} \approx 1 \text{ cm}$ and $\lambda > 10 \text{ km}$. Far more simulations and field measurements should be performed to find an universal formula. However, internal boundary-layers obviously seem to increase z_0^{eff} and the simulations suggested some reasons for this. Local wind profiles were analyzed near shorelines, and they showed that when the flow is from the sea to land, z_0 increased already before the shoreline. In the opposite flow direction decrease in z_0 continued after the shoreline before the sea surface value was reached. The first feature may be related to the formation of an upstream internal boundary layer caused by pressure disturbances while the second may be identified as a wake analogous to those formed after solid obstacles (see also Emeis, 1987). These effects would increase the momentum flux and z_0 in the neighbourhood of rapid roughness changes.

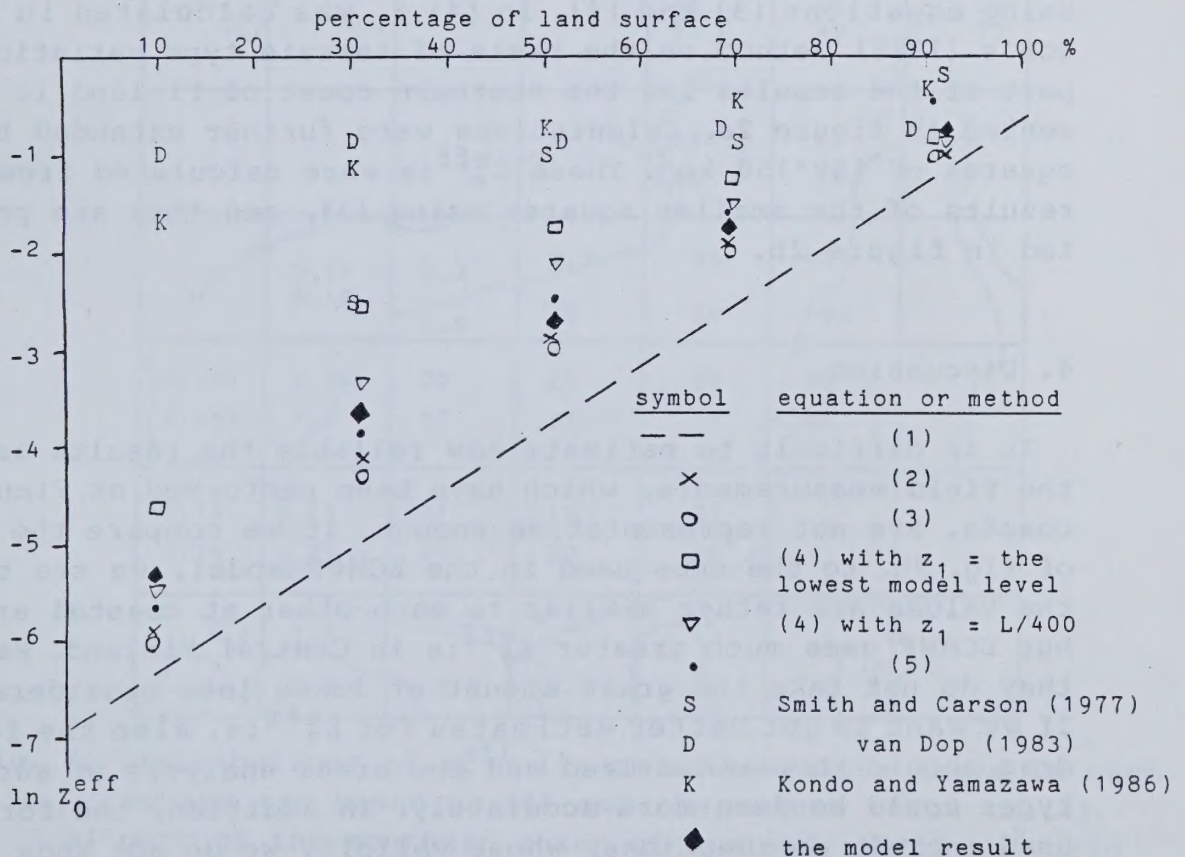


Fig. 1. $\ln z_0^{\text{eff}}$ determined with various methods in the function of land-water-distribution ($[z_0^{\text{eff}}] = \text{m}$)

The effect of the height of the lowest model level was also studied. The lowest level was lifted from 10 to 15 m, and simulations otherwise unchanged were done. However, the resulting z_0^{eff} :s remained exactly the same when determined from the mean flow profile.

3. Calculation of z_0^{eff} for Finland

In practice, when accurate estimates of z_0^{eff} :s are needed for modelling, calculations have been based on terrain classifications. The amount of land and water surfaces and the vegetation types have to be known. Then some of the equations (1) to (5) may be applied, or calculation may be based on some formula developed from the viewpoint of local field measurements. Such calculations have been carried out by Smith and Carson (1977), van Dop (1983) and Kondo and Yamazawa (1986).

The conceptions based on the model simulations were applied to calculation of z_0^{eff} :s for Finland. Surface types were divided into 11 categories and z_0 :s were estimated for each of them, e.g. 0.1 mm for water, 0.1 m for rocky isles and 1 m for forests. Three most common terrain types were picked from each square of $15 \times 15 \text{ km}^2$, and z_0^{eff} was calculated for these squares using equations (3) and (4). In (4) z_1 was calculated in Mason's (1986) method on the basis of terrain type variations. A part of the results for the southern coast of Finland is presented in figure 2a. Calculations were further extended to grid squares of $150 \times 150 \text{ km}^2$. These z_0^{eff} :s were calculated from the results of the smaller squares using (3), and they are presented in figure 2b.

4. Discussion

It is difficult to estimate how reliable the results are as the field measurements, which have been performed at Finnish coasts, are not representative enough. If we compare the z_0^{eff} :s of fig. 2b. to the ones used in the ECMWF model, we see that the values are rather similar to each other at coastal areas, but ECMWF uses much greater z_0^{eff} :s in Central Finland. Perhaps they do not take the great amount of lakes into consideration. If we want to get better estimates for z_0^{eff} :s, also the form drag should be parametrized and the areal analysis of surface types could be done more accurately. In addition, the formulae used include presumptions, whose validity we do not know exactly: the logarithmic profile of the mean flow in (3), and the constant wind speed at height $L/400$ in (4). However, the

results may be somewhat valuable for modelling.

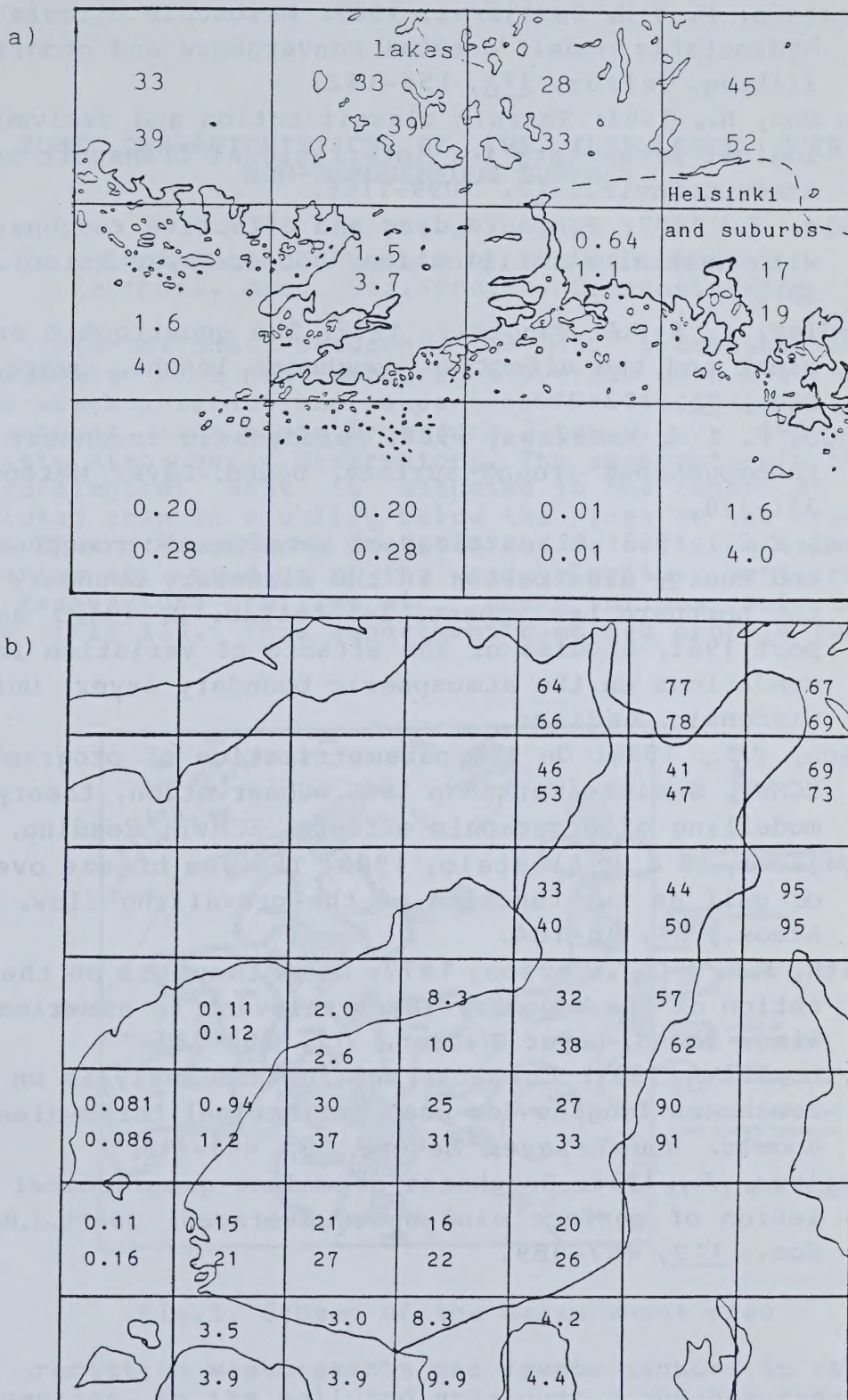


Fig. 2. Distributions of z_o^{eff} (in cm). The values above in each square are based on (3) and the values below on (4).

a) part of the southern coast of Finland, $15 \times 15 \text{ km}^2$ grid squares

b) the whole Finland, $150 \times 150 \text{ km}^2$ grid squares

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SOME CHARACTERISTICS OF THE TURBULENCE OVER NON-HOMOGENEOUS SURFACE

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Interactions between the ABL processes and the influences of a non-homogeneous terrain were the focus of a five-week's international experiment in Czechoslovakia. This experiment took place from June 2 to July 7, 1987 at the Kopisty Atmospheric Observatory. The observatory with an 80m meteorological mast is situated in the centre of a most polluted area in a valley below the ridge of the Krusne hory Mts. Upwind topography profiles can be suitably selected for experimental studies of the wind velocity, vertical wind- and temperature profiles etc. above various real topographies [Fig 1.]. This experiment combined profile turbulence

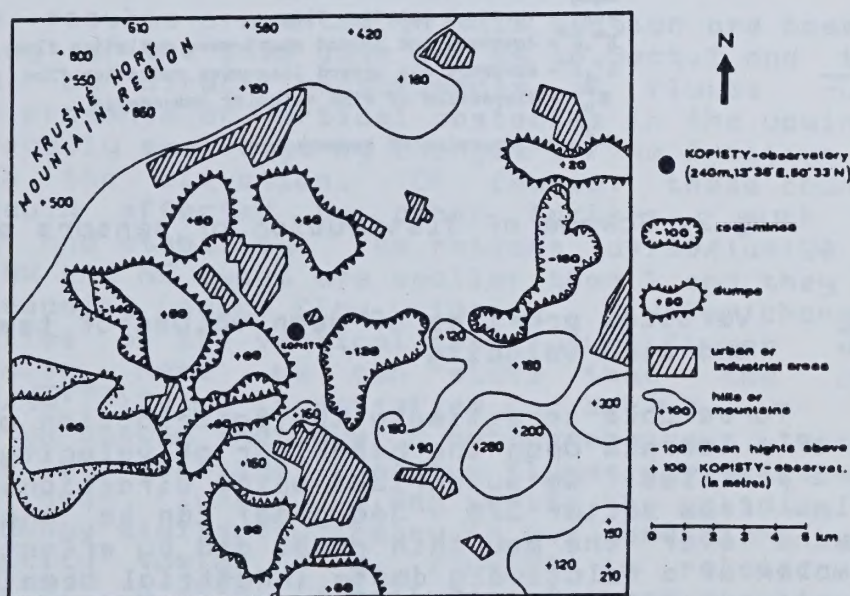


Fig.1. Schema of the measurement area

and radiation measurements and remote sensors to study the kinematics in the polluted area over a non-homogeneous underlying relief. The general results point to the fundamental influence of the non-homogeneous terrain, of the turbulent characteristics affected by it, and of the presence of anthropogenic elements in the lower part of the atmosphere on the formation of the surface layer structure.

The object of this paper is to present some of the

results that analyse the set of empirical data of profile measurements [Fig.2].

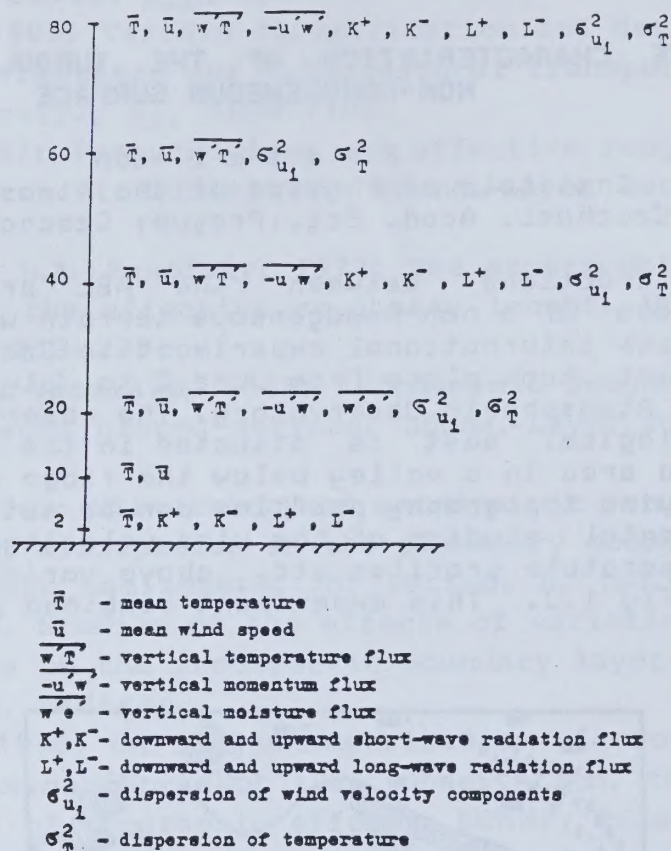


Fig.2. Scheme of distribution of sensors on the mast

Vertical profiles of mean values of temperature and wind velocity

To be able to estimate the contribution of the influence of a terrain upon the behaviour of velocity and temperature profiles, we summarized upwind directions. The NW cases (flow from sector 270 - 360 grad) can be characterized by upwind over the mountain range and by effects of a broad complex of a relatively dense industrial area; on the contrary, a typical feature of the NE cases (010 - 080 grad) is a flow over a relatively not very complicated terrain with not too high scattered land coverage.

To calculate the derivatives $\partial u / \partial z$ and $\partial T / \partial z$ we used an approximation of the measured profiles by the smoothing cubic spline method.

It follows from the analysis of profiles $\partial u / \partial z$ and $\partial T / \partial z$ over both terrain types (NW and NE) that

- there are very significant dependences of the profiles of both quantities upon the type of the underlying relief,

- ideas about a "classic" profile shapes over a

homogeneous terrain cannot be simply transferred to analogous consideration on a non-homogeneous terrain,

- we cannot establish any one-to-one relation between a stability parameter of the z/L type and the vertical temperature gradient $\gamma = -\partial T / \partial z$, the non-uniqueness of this relation being increasing with increasing non-homogeneity of terrain; if it is possible, it is likely to be more advisable to prefer using parameter z/L in these cases,

- an assumption can be made on the existence of critical value of the vertical temperature gradient γ_{crit} (and/or $(z/L)_{crit}$), which is the most favourable for the development of small-scale turbulent eddies; this value will probably be subjected to changes in dependence upon the relief, although not very profoundly,

- the shape of profile $\partial T / \partial z$ especially in the upper part of the surface layer is influenced to a large extent by effects of radiant temperature changes; apart from purely dynamic factors, the real shape of profile $\partial u / \partial z$ is considerably influenced by changes of the temperature profile.

Vertical temperature and momentum fluxes in relation to profiles

The results presented in this section are based on the processing of the same data set as in Sect.1 and they are supplied by direct measurements of fluxes $-u'w'$ and $w'T'$. The presence of vertical obstacles in the upwind stream can be readily seen e.g. by changes of the friction velocity u^* with the elevation. Of course, these changes are considerably affected by other factors, such as flow velocity and stability. The ratios $(u^*)_{80} / (u^*)_{20}$ for both terrain NW and NE types are smaller than 1 and they increase with increasing upwind flow velocity. Similar changes occur in profiles of the vertical temperature fluxes $w'T'$, as well. Also here we can state that the criterion $(w'T')_{80} / (w'T')_{20} \approx 1$ is satisfied for $z/L > 0.1$.

If we assume that over a non-homogeneous terrain the influence of vertical momentum fluxes dominates that of vertical heat fluxes, we can verify the validity of the Monin-Obukhov similarity theory in our conditions by mapping the quantity $(u^*)_{80} \cdot \phi_{M,H} / (u^*)_{20}$ in relation to the parameter z/L . A detailed analysis of these results suggests that it is not possible to employ dimensionless profiles ϕ_M and ϕ_H (frequently called universal function) for the purpose of a further analysis of the vertical profiles of wind velocity and temperature over a non-homogeneous terrain. Under some restrictions, the standard forms can be employed after their modification by introducing in these forms the influence of the dynamic action of the underlying relief.

Some other special properties of the surface layer structure

In the present paper vertical and time changes of selected characteristics are described, e.g. net radiation, vertical temperature fluxes, friction velocities. It follows from the observed energetic characteristics that the most significant influence of terrain concerns the basic turbulent characteristics u^* and $w'T'$. Their changes are then responsible for further changes of microclimatological characteristics of the surface layer.

The results obtained indicate that the vertical temperature fluxes are affected not only by the terrain roughness, but also by the radiant situation in this layer. The presence of atmospheric particles, in particular of anthropogenic origin, has an important effect upon the space and time distribution of these radiant characteristics.

On the Frictional Force in Steady Horizontal Flows,
Tangential Acceleration Included

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EXTENDED ABSTRACT

The objective of this paper is to derive equations for the magnitude and direction of the frictional force in steady horizontal flows, including the non-linear terms for tangential acceleration. The equation of concern reads:

$$\frac{U}{R} \mathbf{k} \times \mathbf{U} + f \mathbf{k} \times (\mathbf{U} - \mathbf{U}_g) = \mathbf{F} \quad (1)$$

Where $\mathbf{U}$ and $\mathbf{U}_g$ are the wind and the geostrophic wind vectors, $\mathbf{F}$ the frictional force vector and the other symbols are standard. From (1) we get that the component F_s of the frictional force along the line of the wind, and the component F_n normal to the wind are given by the equations

$$F_s = -f U_g \sin \beta \quad (2)$$

$$F_n = \left(\frac{U^2}{R} + fU \right) - f U_g \cos \beta,$$

β being the cross-isobar angle. The magnitude F of the frictional force is

$$F = (F_s^2 + F_n^2)^{1/2} = \left\{ f^2 U_a^2 + \frac{U^4}{R^2} - 2f \frac{U^2}{R} (U_g \cos \beta - U) \right\}^{1/2}, \quad (3)$$

U_a being the ageostrophic wind. Its magnitude is obtained from the triangle of velocities $\mathbf{U}$, $\mathbf{U}_g$ and $\mathbf{U}_a$ in Fig 1:

$$U_a^2 = U^2 + U_g^2 - 2UU_g \cos \beta. \quad (4)$$

As to the direction of the frictional force, the angle $\pi + \alpha$, $\text{tg}(\pi + \alpha) = \text{tg} \alpha$, between the wind and the frictional force vectors is

$$\text{tg} \alpha = \frac{F_n}{F_s} = \cot \beta - \frac{\frac{U^2}{R} + fU}{fU_g \sin \beta}. \quad (5)$$

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By taking the algebraic sum of the normal components of the four forces involved (pressure gradient, Coriolis, centrifugal and friction), and equating the sum with zero, we get as the solution of the equation the frictional gradient wind, viz.

$$U = \frac{fR}{2} \left\{ -1 + \left[1 + \frac{4}{fR^2} (fU_g \cos \beta - F \sin \alpha) \right]^{1/2} \right\}, \quad (6)$$

which for $F = 0$ reduces to the well-known gradient wind equation, namely

$$U_g = \frac{fR}{2} \left[-1 + \left(1 + \frac{4U_g}{fR} \right)^{1/2} \right]. \quad (7)$$

If $|R|$ is very large (= curvature of streamlines negligible), we arrive at the symmetrical relationships

$$P = fU_g, \quad C = fU \quad \text{and} \quad F = fU_a, \quad (8)$$

P being the magnitude of the horizontal pressure-gradient force and C that of the Coriolis force: We show further that in these flows (negligible centrifugal force) the following simple relationship holds among the three relevant angles of the flow:

$$\alpha + \beta + \gamma = \frac{\pi}{2} \quad (9)$$

The meaning of the angle γ is shown by Fig. 1.

We mention for the sake of comparison that in the Ekman spiral, for which the constancy of the geostrophic wind with height is assumed, the surface wind U_0 is given by,

$$U_0 = U_g (\cos \beta - \sin \beta) \quad (10)$$

whereas from (6) we have that

$$U_0 = U_{g0} \cos \beta_0 - U_{a0} \sin \alpha_0. \quad (11)$$

An example will show how accurately (11) predicts the surface wind for a well-known wind profile.

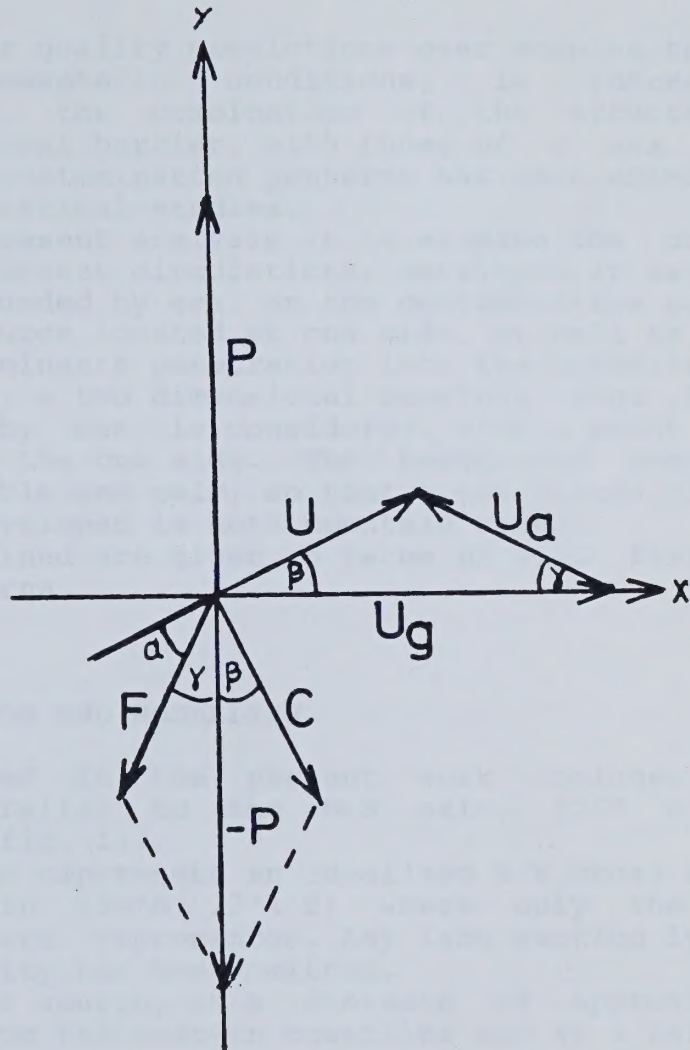


Fig. 1. Diagram of the forces and velocities for the lower 100-m layer of the Leipzig Wind Profile.

AN ANALYTICAL STUDY OF NATURAL BARRIER EFFECTS ON SEA BREEZE AND DISPERSION

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1. INTRODUCTION

The need for air quality predictions over complex terrain, under various atmospheric conditions, is increasingly recognized. Though, the combination of the effects of a relatively high natural barrier, with those of a sea breeze circulation on the contamination patterns has been examined in rather limited theoretical studies.

The aim of the present analysis is to examine the combined effect of the sea breeze circulations, developed at each side of a mountain surrounded by sea, on the contamination patterns due to a point source located at one side, as well as on the degree of the contaminants penetration into the opposite side.

For this purpose, a two dimensional mountain range, 1000 m high, surrounded by sea is considered, with a point source 40 km offshore from the one side. The background conditions are considered stable and calm, so that a sea breeze circulation can be well developed in both mountain sides.

The results obtained are given in terms of wind field and concentration patterns.

2. PROBLEM DEFINITION AND MODELLING

The terrain used in the present work includes a 2-D mountain range, parallel to the N-S axis, 1000 m high, surrounded by sea (fig. 1).

The above terrain represents an idealized E-W cross section of the Athens basin (38°N, 23°4'E) where only the gross terrain features are represented. Any land section lying in the mountain proximity has been omitted.

A point pollutant source, at a distance of approximately 40 km offshore from the eastern coastline and at a height of 20 m, is assumed.

The present analysis has been made using a two dimensional domain, covering 192 Km horizontally and 6 Km vertically. The grid consists of 45x35 mesh points. The domain discretization around the mountain is shown in fig. 1.

The initial atmospheric conditions are described by a near zero geostrophic wind ($u_g = 0.4$ m/s) and a stable temperature profil as shown in fig. 2. The initial release from the point source is taken equal to 10^6 units/sec. The initial concentration pattern is shown in fig. 3.

The model utilizes mass, momentum, energy and pollutant concentration balance equations. In the present analysis the integrated concentrations along the lateral y direction are calculated. The above equations are discussed in detail in the ADREA-I code description /1,2/ utilized for the present

application.

For the eddy viscosities modelling the k-l concept has been adopted as developed in ref. /3/. The analytical expressions governing the turbulent diffusion modelling are given in refs. /3,4,5/.

The ground surface temperature T is assumed to vary diurnally according to a sinusoidal function of time, yielding a maximum value at 14:00 and a minimum at 02:00 LST:

$$T(t) = T(t_0) + 12\sin\left(\frac{2\pi}{86400} (t-t_0+79200)\right) \quad (2.1)$$

where t = time from local midnight in seconds
and $t_0 = 21600$ sec.

The calculations start at 06:00 LST assuming that the horizontal gradients of all variables, except pressure, are equal to zero.

3. RESULTS

The calculated flow fields and concentration patterns (the concentration is normalized by its maximum value) are given at 11:00, 15:00 and 19:00 LST, as shown in figs. 4 to 13.

A generation of a sea breeze circulation is already distinguished at both sides of the mountain by 11:00 LST (fig. 4). The combination of the inland advancing sea breezes and the anabatic winds has led to an intense upward motion around and over the top of the mountain.

The surface flow of the east side circulation advects highly contaminated air masses to the east foot and slope of the mountain while lower concentration fractions have already penetrated into the ascending air masses and the seaward currents of both circulations (fig. 5).

The sea breeze circulations attain their maximum intensity at 15:00 LST. The strong upward motion created around and over the mountain top (figs. 6, 10) lead to a quite high boundary layer (fig. 11). By this time, air masses of high concentrations have been spread along the east slope and over the mountain top, slipping into the upward current and the offshore flows of the sea breezes (fig. 7). The return flow of the west side circulation is obviously responsible for the contaminants spreading in the upper atmospheric layers of the west section (see also fig. 12). In low altitudes the mountain presence seems to exert a drastic control on the concentration distribution, since the pollutant concentration fractions at the west side below the mountain top, are by five to six orders of magnitude lower than the east side values.

By 19:00 LST the circulations have considerably subsided (fig. 8) leading to a reduction of the concentration fraction along the east slope by one order of magnitude (fig. 9). However, air masses of considerable contamination are maintained there, as well as in the region over the landward currents of the circulations.

4. CONCLUSIONS

In case of stable and calm atmospheric background conditions, over a terrain consisting of a 2-D mountain range, 1000 m high, surrounded by sea, with a point source located at 40 Km offshore from one side, the flow and contamination patterns as well as the mixing layer depth evolve as follows:

1. At each side of the mountain a sea breeze circulation is formed. The inland advancing landward currents are reinforced by the upslope mountain winds, resulting to a strong upward motion around and over the mountain top, so that a quite high boundary layer is developed.

2. The contaminants released from the point source are mainly advected in high altitudes due to the upward winds where they penetrate into the two seaward currents of the circulations.

3. In the opposite side of the mountain, the pollutant concentration distribution is mainly affected in the upper layers, due to the offshore circulation current. In the lower altitudes the mountain presence exerts a drastic control on the concentration distribution, keeping the opposite side relatively clear.

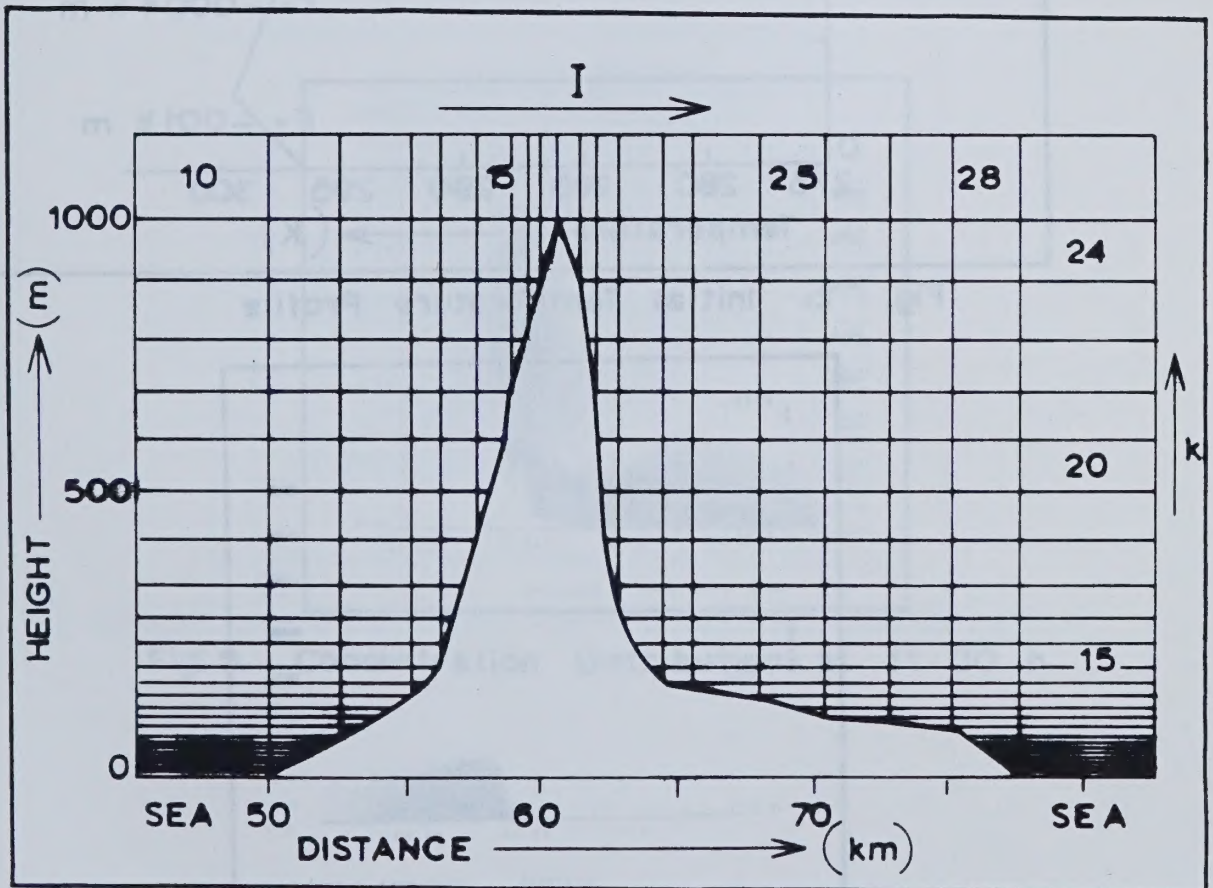


Fig. 1. The Domain Discretization around the Mountain

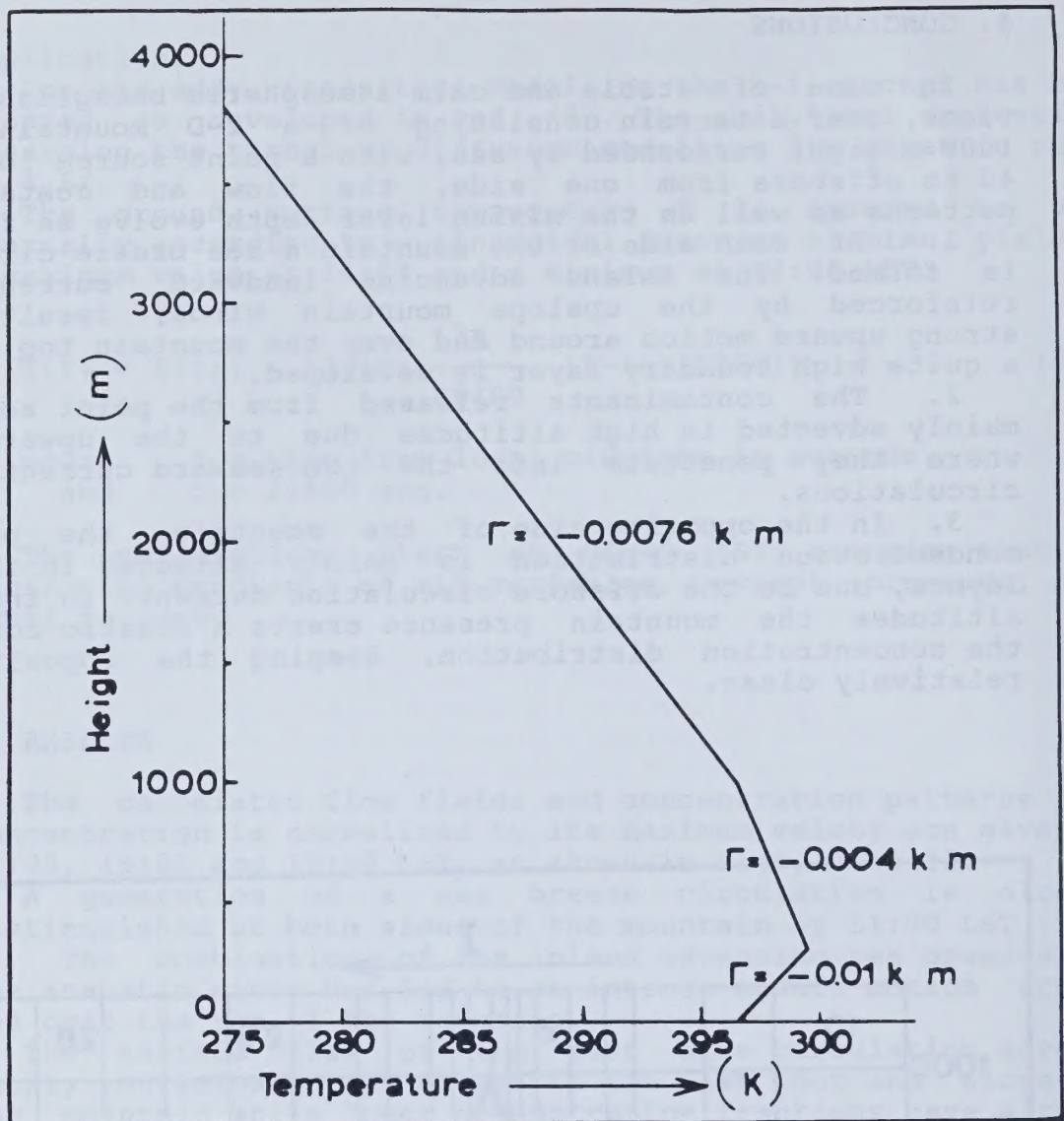


Fig. 2. Initial Temperature Profile

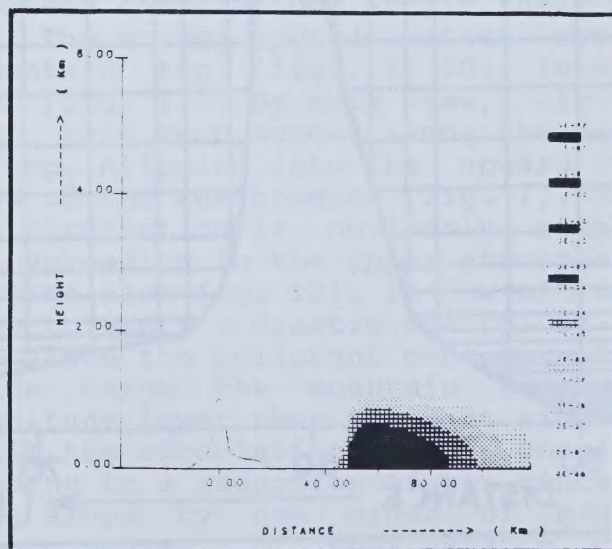


Fig. 3. Initial Concentration Distribution (06:00 h)

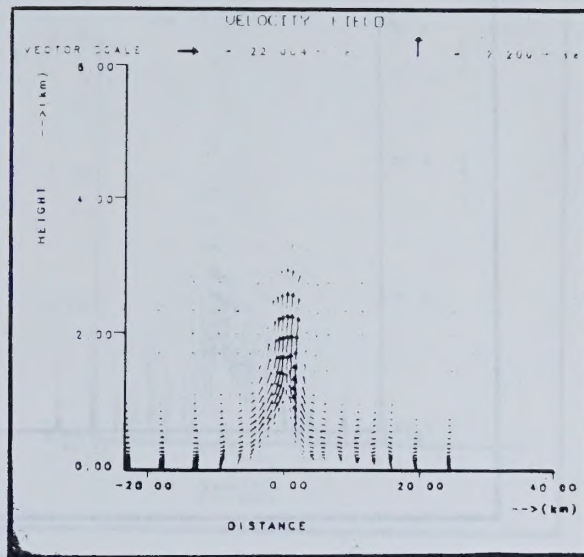


Fig. 4. Flow Field at 11:00 h

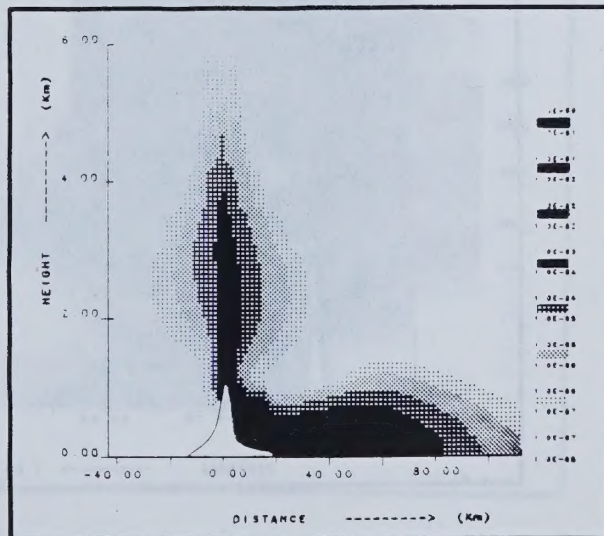


Fig. 5. Concentration Distribution at 11:00 h

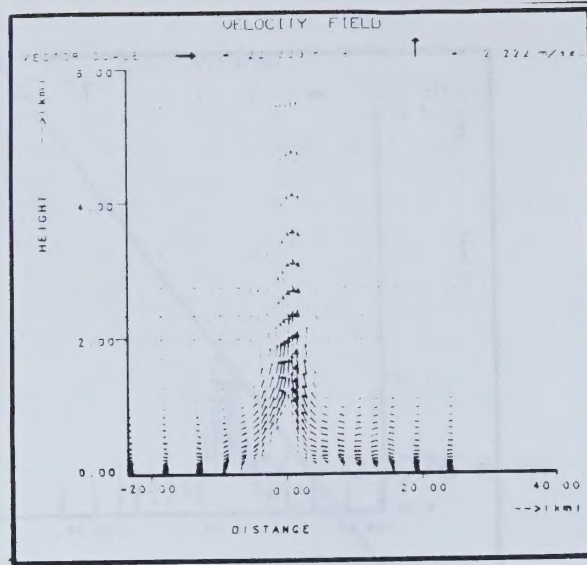


Fig. 6. Flow Field at 15:00 h

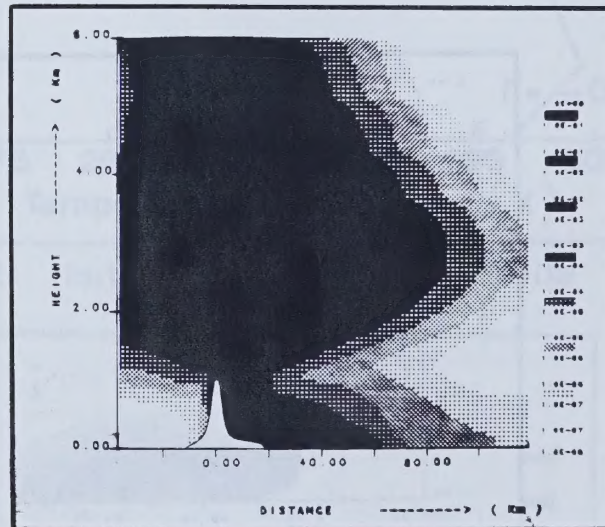


Fig. 7. Concentration Distribution at 15:00 h

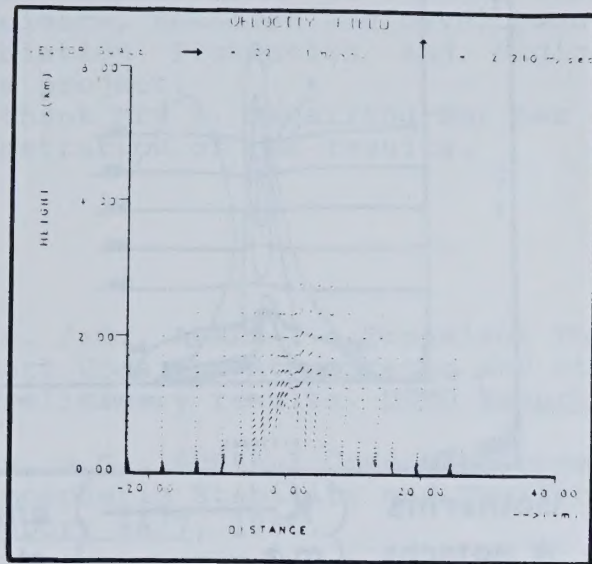


Fig. 8. Flow Field at 19:00 h

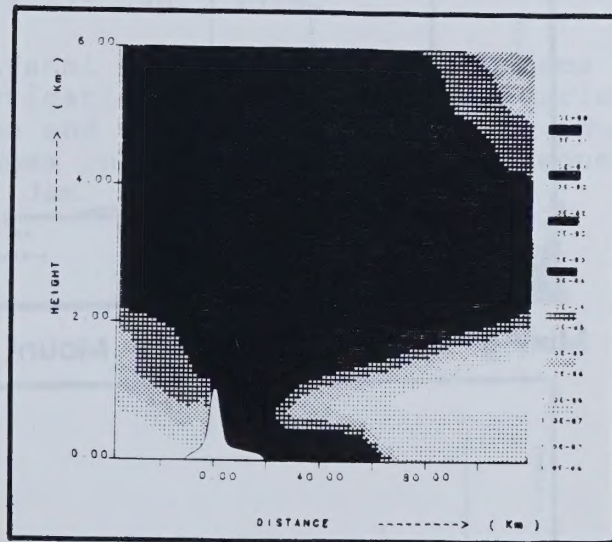


Fig. 9. Concentration Distribution at 19:00 h

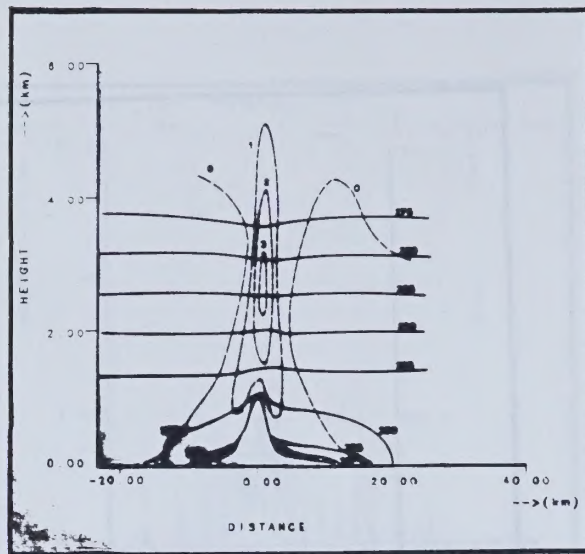


Fig. 10. Isotherms $\left(K \text{ —————} \right)$ and
W Isotachs $\left(m/s \text{ - - - - -} \right)$ at 15:00 h

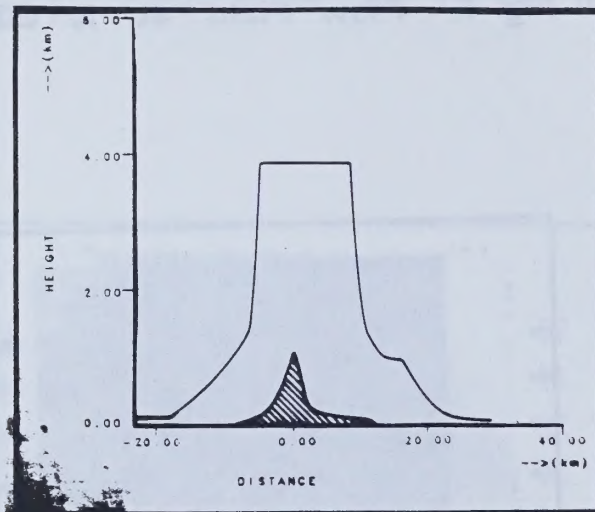


Fig. 11. Mixing Depth around the Mount at 15:00 h

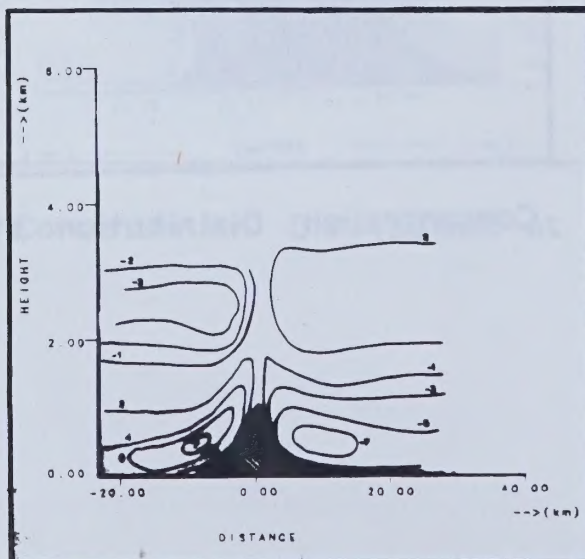


Fig. 12. u Isotachs at 15:00 h

ACKNOWLEDGEMENTS

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AIR FLOW ANALYSIS ON THE WINDWARD SIDE OF THE DINARIC ALPS DURING THE BORA OF MARCH 6, 1982

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Abstract

Analysis of the bora in the period from March 5 to March 8, 1982, is presented. The bora started at almost the same time along the whole Northern Adriatic coast. The maximum wind speed was recorded on March 6, 1982. Aircraft measurements showed high wind speed in the lower troposphere, resulting in high values of the Froude number. In the basins of Slovenia, relatively strong winds were recorded as well. Surface potential temperature analysis showed minimal differences among various locations, indicating no stagnation in the flow before the mountain crest.

Introduction

On March 4, a cold air invasion, coming from NW, reached the Alps. The major part of the cold air turned to the South towards the Mediterranean sea, forming a secondary cyclone over the Gulf of Genoa. The rest of the cold air moved towards the Adriatic around the Alps on the Eastern side. While passing the Karst and Dinaric Alps it caused a strong bora. The first bora gust was recorded near Ajdovščina (Karst region) in the night on March 5, 1982 at 00:00 GMT. The beginning of the bora was violent and the velocity of the bora gusts quickly exceeded 20 m/s. Within two hours the bora was recorded along all the Northern Adriatic coast. The bora strengthened next day, when its gusts reached 51.5 m/s, the highest bora wind speed of 1982. On March 6, ALPEX aircraft measurements of the bora were made by an NCAR Electra research aircraft. In the next two days the bora force decreased and from March 8 to March 9 terminated. The bora was carefully analyzed on March 6, using aircraft data together with surface observation data.

Three Layer Hydraulic Model of the Bora

Hydraulic theory was often applied in bora research after the ALPEX research experiment. (Smith, 1984, Urbančič, 1988a and 1988b). The bora can be described by a

hydraulic model in which the main processes are found in three significant layers:

- the upper layer
- the interaction layer (transition layer)
- the lower layer (the bora layer)

The upper layer has quite a uniform air flow which is weakly (or even not) influenced by orography. Its direction is statistically insignificant, but it seems to be in a SW direction most frequently. (Figure 1)

The lower layer is found as a cold air flow blowing from the continent to the Adriatic. Its potential temperature can be over 10 K lower than in the upper layer. Approaching the coastal mountain crest it is accelerated. The greatest acceleration is found in the lee side of the mountain crest where the maximum velocity is reached. On the ground this wind is called the bora and it blows along the northern Adriatic coast. Orography is the most important factor determining its characteristic NE or E direction.

The interaction layer lies between the upper and lower layers. It is characterized by interaction and mixing of the two air flows of different temperature, direction and speed. The wind speed in this layer normally decreases with height, while the wind direction changes from the direction of the lower to the direction of the upper flow. The upper layer is potentially warmer and the interaction layer has great static stability.

The interaction between the two major layer is not significant on the windward side of a mountain crest where the interaction layer is thin relative to the lee side, where a hydraulic jump is expected.

The paper is concerned with the flow of cold air in the lower layer of the three layer model. The distribution of surface potential temperature and Froude number can be used to determine the type of bora flow.

Analysis of ground measurements

The ground potential temperature at 20 meteorological stations (Figure 2) was quite uniform along the expected trajectories of the cold air flow. The first expected trajectory: Zagreb - Karlovac - Reka (stations No. 7, 8, 9, 11, 13, 14 and 2) has an approximately 2 K higher potential temperature than the second one: Sisak - Plaški - Senj (stations No. 10, 16 and 1). The data from Lika (stations No. 18, 19 and 20) that lies on the windward side of the Velebit, close to the second expected trajectory of the cold air flow, show good agreement with the data corresponding to that trajectory. In this area the only higher potential temperature is recorded in Zavižan (station No. 17) which is situated at the top of the Velebit about 1700 m above sea level. This point is not a representative ground potential temperature. The value of 279 K seems to be the characteristic ground potential temperature along the trajectory of the bora wind in the Senj region.

During an outbreak of bora there is sometimes no surface wind on the windward side of the Dinaric Alps. In this particular case high wind velocities were also recorded at meteorological stations located in the valleys and basins in Slovenia. An Eastern wind of maximum velocity 15 m/s and average velocity 7 m/s in Novo mesto and only slightly weaker winds were observed in Ljubljana and Celje. Winds with over 5 m/s seldom occur at the above mentioned locations.

Computation of Froude Number

The following definition of Froude number is used:

$$Fr = \frac{U}{\sqrt{g^* L}} \quad (1)$$

where U and L are the characteristic velocity and depth of the flow, and g^* is the buoyancy parameter ($g^* = g \frac{\delta \rho}{\rho}$)

The results of the computation of the Froude number are presented in Figure 3. Each curve in the figure shows the results of a computation using measurements at one flight level. The parameters of the Froude number are obtained in the following way:

- The characteristic velocity U is taken as the observed average velocity in the flight level. If there is greater velocity in any one flight level below, the maximum average velocity from the level below is chosen.
- The depth of the flow L is represented by the height of the flight level above the ground.
- The static stability was used in computation of the buoyancy parameter g^* . It was determined from the measured surface potential temperature and the potential temperature at the flight level.

The results of the computation can be used as an estimation of the Froude number only if the data were taken from the top of the lower layer and the beginning of the interaction layer. The dashed line in Figure 3 denotes the most probable value of the Froude number.

On the windward side of the mountain crest (right boundary in Figure 3) the interaction layer lies between the second and the third flight level (Smith, 1987, Urbančič, 1988a). This means the physically acceptable value of the Froude number on the windward side distant from the crest should be between 0.7 and 0.95. This would be in agreement with computations of Froude number based on Zagreb radiosondings. Smith (1984) reported a value of 0.62 and Vučetić (1985) reported 0.74.

Approaching the mountain crest the cold air flow becomes thinner. Just above the ridge the interaction layer lies between the first and the second flight level and the value

of the Froude number should be between 1.2 and 1.9 . The dashed line in Figure 3 denotes the most probable values of the Froude number.

On the lee side an intensive drop of Froude number is evident. Unfortunately, immediately after passing the ridge, the lower layer sank below the first flight level, so the Froude number cannot be determined. Urbančič (1988a) reports the appearance of a hydraulic jump there which causes a large change in Froude number. At the left boundary where the lower layer originates, the data from the first flight level can be used to compute the Froude number, which probably lies between 0.4 and 0.5 .

Conclusion

The results of the computation show that the Froude number is greater than 1 on the windward side of the mountain crest and that it drops far below 1 on the lee side.

The aircraft measurements also show a violent turbulence immediately after passing the mountain crest that could be explained by a hydraulic jump (Smith, 1984,1987, Urbančič 1988a). This is in agreement with the demonstrated drop of Froude number on the lee side of the mountain crest.

The uniform ground potential temperature which was found on the windward side shows either that the lower layer has no potential temperature gradient or that there is no hold-up of the cold air flow before the mountain crest. In both cases a high value of the Froude number is expected.

The violent beginning of the bora and high wind velocities in the basins on the windward side of the Dinaric Alps confirms the supposition of supercritical flow. According to the classification of air flows over a mountain crest based on hydraulic theory (Urbančič,1988a,1988b), this bora event can be classified as a type B flow.

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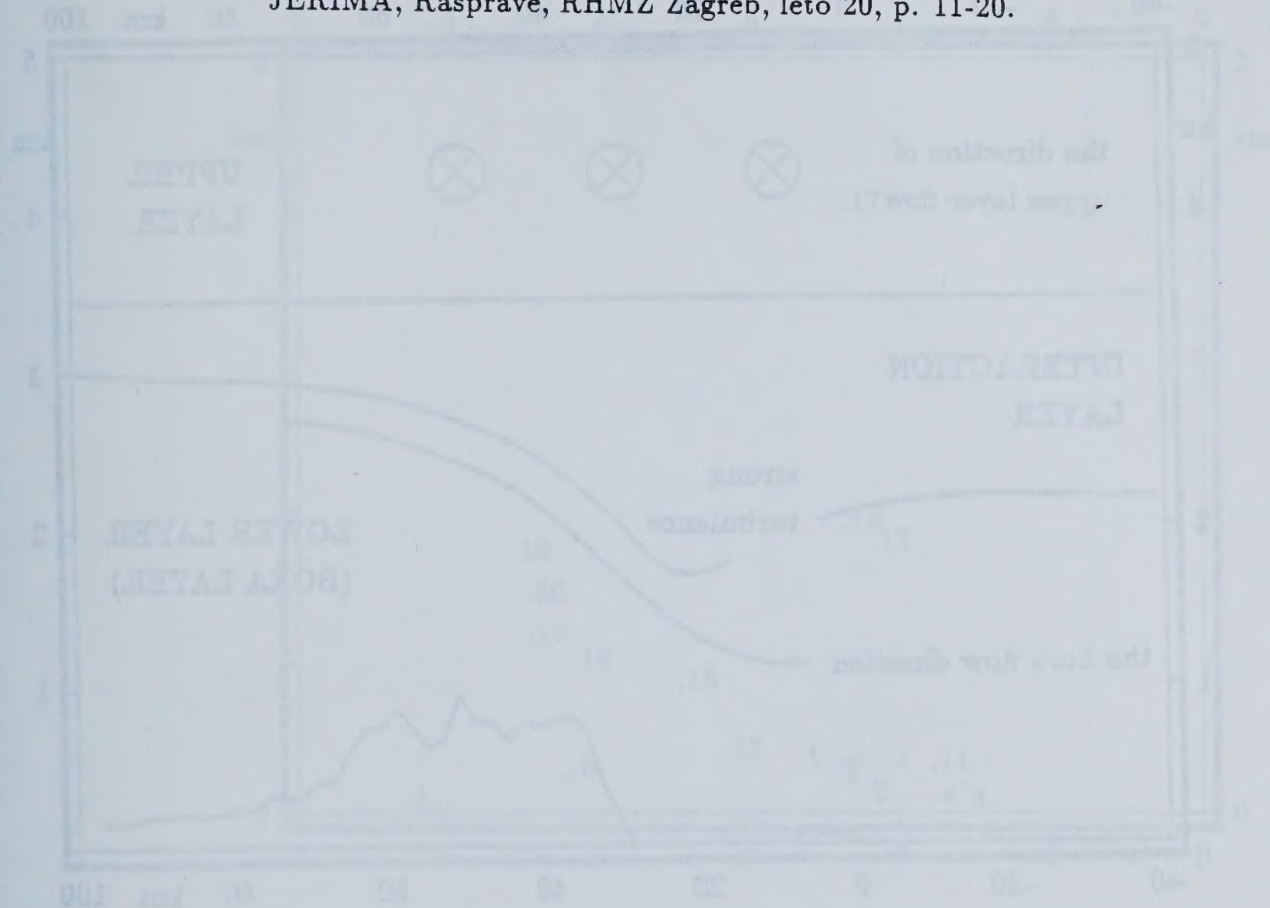


Figure 2: The dependence of ground surface temperature on March 9, 1987, at 2 km. The ground surface temperature was measured at 20 stations. The meteorological stations are: 1. Zagreb, 2. Ljubljana, 3. Trieste, 4. Udine, 5. Gorizia, 6. Trieste, 7. Trieste, 8. Trieste, 9. Trieste, 10. Trieste, 11. Trieste, 12. Trieste, 13. Trieste, 14. Trieste, 15. Trieste, 16. Trieste, 17. Trieste, 18. Trieste, 19. Trieste, 20. Trieste.

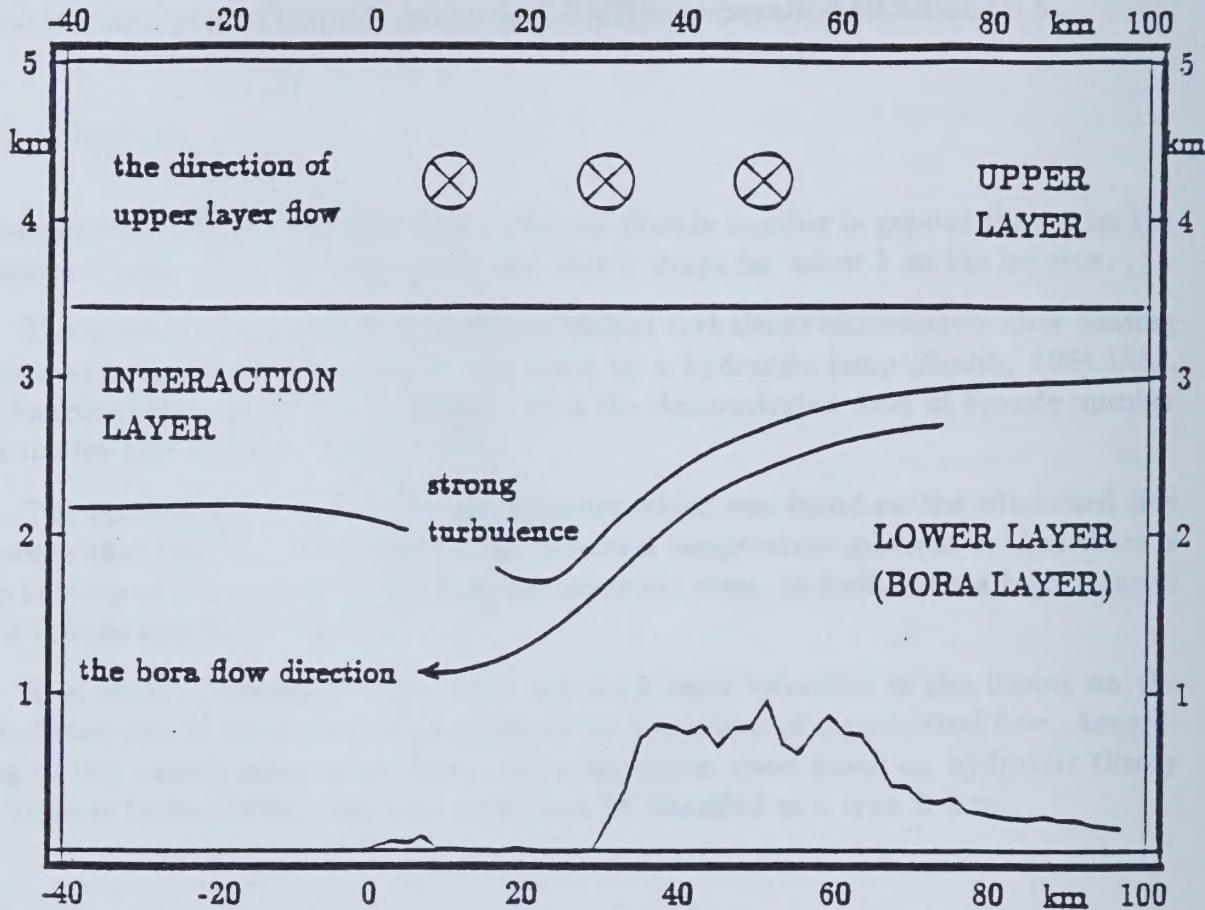


Figure 1: Scheme of three layer hydraulic model of the bora

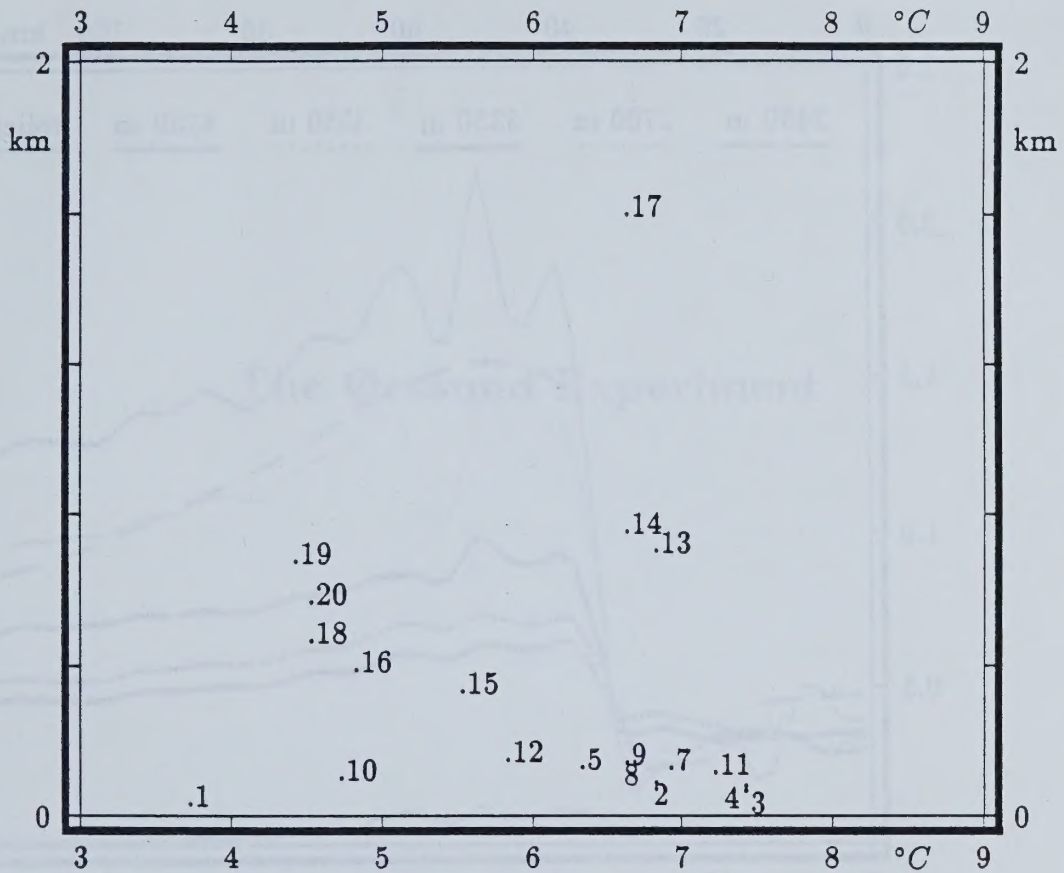


Figure 2: The dependence of ground potential temperature on March 6, 1982, at 2 pm. on height above sea level (km). Measurements from the following meteorological stations: 1. Senj, 2. Letališće Reka, 3. Letališće Pulj, 4. Šibenik, 5. Split - Marjan, 7. Zagreb - Maksimir, 8. Jastrebarsko, 9. Piskarovina, 10. Sisak, 11. Karlovac, 12. Vojnić, 13. Delnice, 14. Vrelo Ličanke, 15. Ogulin, 16. Plaški, 17. Zavižan, 18. Ličko Lešće, 19. Titova Korenica, 20. Gospić

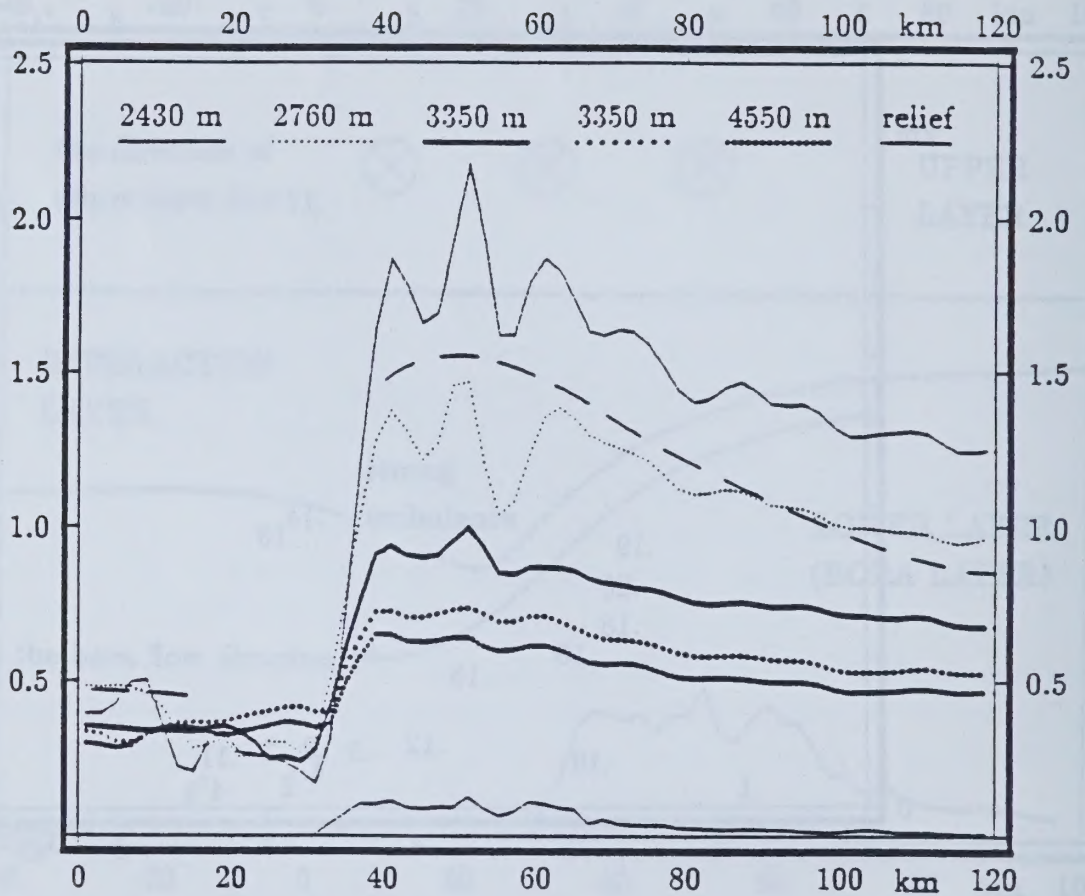


Figure 3: Results of computation of Froude number under the following assumptions: flow depth is equal to the flight level, flow velocity is equal to the maximum velocity measured at the flight levels below or at a certain level, and 279 K is taken as the ground potential temperature. The dashed line denotes the most probable value of Froude number.

The Øresund Experiment

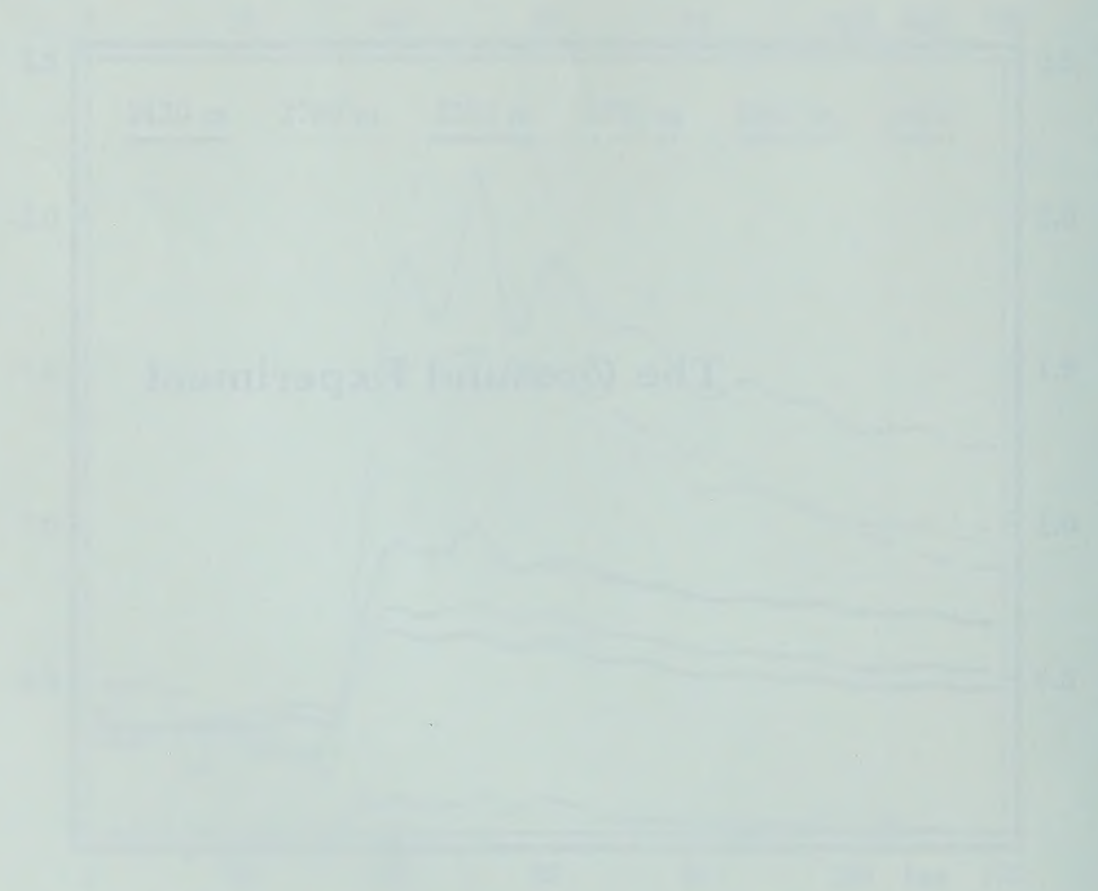


Figure 1. The Normal Experiment. The graph shows the results of the experiment. The y-axis represents the percentage of the total, and the x-axis represents time. The four lines represent the total, wet, dry, and the difference between the total and the wet and dry components. The total line starts at 0.8, peaks at 0.95 around time 20, and then fluctuates between 0.8 and 1.0. The wet line starts at 0.6, peaks at 0.8 around time 20, and then fluctuates between 0.6 and 0.8. The dry line starts at 0.4, peaks at 0.6 around time 20, and then fluctuates between 0.4 and 0.6. The difference line starts at 0.2, peaks at 0.4 around time 20, and then fluctuates between 0.2 and 0.4.

The Öresund experiment - a short introduction

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Background and aim of the experiment

The idea to carry out a mesoscale experiment somewhere in the Nordic area and in cooperation between groups of scientists from Denmark, Finland, Norway and Sweden was born about ten years ago in connection with a joint project called 'Mesoscale Dispersion Models'. This project was an effort to co-ordinate ongoing modelling work in the various Nordic countries and was made possible by the organization Nordforsk. The aim of the proposed experiment was to put together expertise and equipment available in the Nordic countries in order to obtain a high quality data set to test meteorological and dispersion models on the meso-scale. For various reasons it took several years before the idea could be realized, but around 1982 the detailed planning for an experiment to be carried out in 1984 started, again with Nordforsk as the umbrella organization.

The reasons for choosing the Öresund area

One of the obstacles in the early discussions for the realization of a joint Nordic mesoscale field experiment was the choice of a particular geographical area of interest for all parties involved. It was clear, however, that all the countries had an interest to study conditions related to coastal areas, because in all the Nordic countries population centre and industry are to a large extent concentrated to such environment. The Öresund area was finally considered a good choice, because

- Interesting geographical features: two, relatively flat land areas separated by a strait, ca. 20 km wide. In particular during spring and early summer large thermal contrasts between land and water surfaces are likely to cause important mesoscale modification to the flow field in the area.
- The area is densely populated with many industries on both sides of the strait
- Two of the Nordic countries, Denmark and Sweden are directly involved.

General outline of experiment plan

In a series of planning meetings with participants representing all the interested research groups a general outline of the experiment plan was worked out. For the realization of this plan a small 'general staff' headed by Sven-Erik Gryning from Risö was appointed. The plan can be summarized accordingly:

- The experiment should be carried out in a one month period in early summer.
- During this period a week of *intensive measurements* was planned to take place; during this period we expected participation of the Finnish research vessel Aranda and an instrumented airplane from NILU

- Measurements in several towers and with acoustic sounders (sodars) continuously; dense radio sonde ascents at two sites - one on the Swedish side of Öresund and the other on the Danish side
- Tracer experiments (SF_6) across the Öresund as often as possible.

The realization of the experiment

The time period 16 May to 14 June 1984 was selected for the experiment. Many research groups from all the Nordic countries (Denmark, Finland, Norway and Sweden) participated actively as well as groups from Belgium, Holland and West Germany. The experimental plan outlined in the previous paragraph was executed successfully in every respect.

Results so far

- A very complete data base, with free access for the scientific community has been prepared
- Results of analyses of the data, including tests of the performance of various meteorological and dispersion models for the mesoscale were presented at two workshops:
Workshop I in Copenhagen, October 1986
Workshop II in Uppsala, October 1987.

Results of the Øresund Experiment - A Resumé of Previous Workshops

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1 INTRODUCTION

In 1984 the Øresund experiment was carried out with the purpose of investigating the modifications in the wind field and the atmospheric dispersion over a region with cold water and warm land surfaces. One of the goals of the experiment was to obtain a data set that can be used for verification of mesoscale models in a coastal environment. Gryning (1985) gives a detailed description of the project.

The experimental campaign was carried out over the Øresund, the strait between Denmark and Sweden. The measuring campaign took place during the period 16 May to 14 June, 1984. During 4 - 10 June a special intensive measuring programme was carried out. From a scientific point of view the Øresund region is well suited for an investigation of this type as the coastlines on both sides of the strait are rather straight and practically parallel, and the land area is fairly flat. The width of the strait is about 20 km. A data bank containing all the measurements has been established and is available for general use.

The data bank has been in great international demand. The data evaluation has not only been carried out by groups from the Nordic countries, but groups from all over the world have participated.

Results from the preliminary analysis of the data were presented at the workshops in October 1986 and October 1987, respectively. Soon it became apparent that intriguing physical mechanisms were at play on the various days of the experiment.

2 METEOROLOGY

The meteorological analysis soon concentrated on the effect of the cold water surface on the wind field.

Doran and Gryning (1987), Gryning and Doran (1987) modelled the wind field over the Øresund area at easterly flows using a modified version of Pielke's model.

The simulated behaviour of the winds is complicated, particularly at low levels. Close to the Swedish coast, the stable layer over water still was relatively shallow and weak, and the offshore winds in the lowest 200 m increased in response to the lower surface roughness of the water. However, an examination of the turbulent exchange profiles farther from the coast showed that momentum transport from above was inhibited in the strongly stable layer near the sea surface. Eventually, the frictional drag of the surface caused the near-surface winds to decrease even though the roughness length

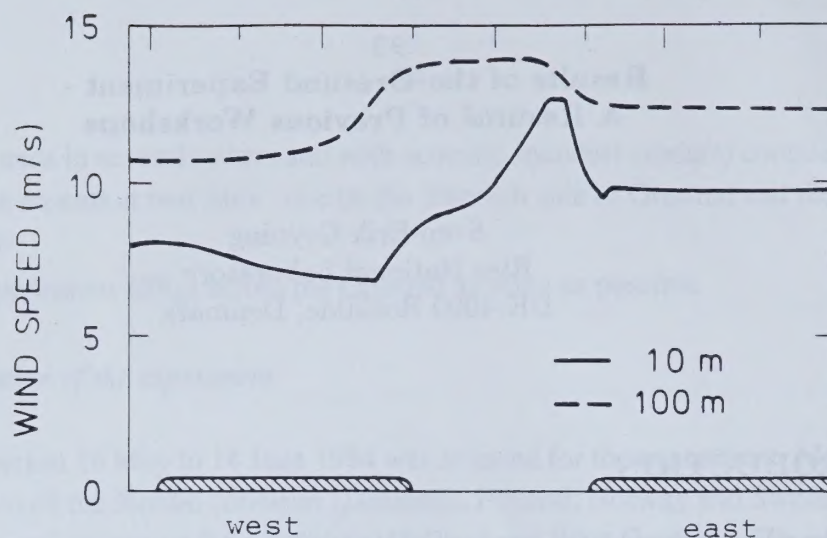


Figure 1: Simulated horizontal profiles of wind speed at 10 and 100 m at 1200 CET on 5 June, 1984.

remained small. As the wind encountered the second heated land mass, the increased mixing again produced acceleration, despite the substantially larger roughness length. At higher levels, however, the winds were decoupled from the surface by the shallow stable layer. This resulted in a lower drag on the winds aloft, and they accelerated over the water. When land was encountered after passage over water, the well-mixed atmosphere responded to the larger surface roughness, and the upper-level winds decelerated. This behaviour is illustrated in Fig. 1.

At a conference in 1987 it was suggested that the effect was due to blocking and not stability as suggested above. This led Doran (1987) to repeat the simulation, replacing Copenhagen by water. A similar behaviour of the wind field over water was found which excluded the blocking hypothesis. It has also been suggested that the effect is caused by the induced pressure gradient because the air over the water near Copenhagen is colder than near Sweden (easterly wind). This hypothesis has not yet been tested.

Gryning and Joffre (1987) analysed the measured wind speeds. This analysis is confined to data from the intensive period of the Øresund experiment only (3–10 June) in which the wind was persistently easterly. Data were used from the small mast on the Barsebäck coastline, from the research vessel Aranda, and from the mast at Charlottenlund. The second part of the data analysis deals with the wind at the 50-m level and above. This analysis is based on the wind measurements performed by the sodars.

Because of easterly wind throughout the entire period, the measurements at Barsebäck represent the conditions where the water surface has not yet imposed its effect on the flow. It was clearly seen that for water fetches up to 10–15 km the wind speed was higher over the water. Further downwind the wind speed decreased, and for water fetches larger than 20 km, the wind speed was always smaller than at Barsebäck. The ratio of the wind speed over water to the upwind wind speed at the 10-m level was analysed. The points were found to order along a line when plotted as a function of a dimensionless distance:

$$X = \frac{g}{T} x (T_{Bb} - T_w) / u_{sea}^2$$

where g/T is the buoyancy parameter, x is the water fetch, T_{Bb} the air temperature at Barsebäck (land condition), T_w is the temperature, and u_{sea} the wind speed over the water. The dimensionless distance can be regarded as a lateral Richardson number.

Andrén (1987) has carried out three-dimensional simulations with a higher-order closure model for the conditions of easterly flows. The model was first run for daytime conditions on 4 and 5 June with $31 \times 31 \times 16$ grid points covering the major parts of Skåne and Sjælland. A preliminary evaluation of the runs focused on the development of the boundary layers. The model boundary layer height over Skåne grew to a value of about 1000 m at noon both days, whereas the radiosonde potential temperature profiles clearly showed that the boundary layer never exceeded 500 m. Of course, this also affected the inland deepening of the modelled internal boundary layer on the Danish side. It was suggested that this was caused by the model domain. The eastern lateral boundary was west of the Skåne eastern shore line. Thus the model saw an infinite land area to the east and did not feel the rather cold water in the the southern Baltic.

As the air was advected over the cold Øresund water, stratification in the lower layer immediately switched from convective to stable. A shallow layer with decreasing wind developed downstream. When compared with measurements from the research vessel Aranda it was found that predicted winds were too low in the lower ten metres. The reason for this is likely to be found in the length scale formulation for stable stratification where the depth of this stable layer is given according to a steady state formula. The values for the depth of the stable boundary layer were too large immediately after the shore line. The deepening internal boundary layer was reasonably predicted inland of the Danish Øresund coast considering the simplifications made in the closure scheme.

An analysis of the pressure field on June 4 and 5 (two heavily modelled experimental days) was carried out by Melas (1987). The general meteorological conditions (wind speed, stability, and wind direction) were quite similar these two days except that on 4 June the tracer concentrations were twice those measured on 5 June. General agreement exists between the calculated surface geostrophic and the measured wind on 4 June, but large differences were found on 5 June. This could be caused by baroclinity.

The aircraft measurements of the boundary layer height over Copenhagen were used to evaluate a model developed by Batchvarova and Yordanov (1986), and good agreement was found between measurements and predictions (Batchvarova and Gryning, 1987).

3 TRACER SIMULATIONS

Simulations of the dispersion experiments were carried out with the Matthew/Adpic emergency response model from Lawrence Livermore National Laboratory, puff models, and Gaussian trajectory models. Generally speaking, and contrary to the meteorology, no firm conclusions have been drawn from the preliminary studies. The results are confusing and difficult to understand.

Simulations with the Matthew/Adpic code (Gryning and Gudiksen (1987); Gudiksen and Gryning (1987)) overestimated the measured tracer concentrations on all experimental days except on 5 June. The general overprediction of the SF_6 concentrations by the Matthew/Adpic models probably reflects meteorological processes not modelled by the code. The SF_6 , released into the transition layer, was initially subjected to a slight downward motion as it was transported across the upwind coastline (see Figs. 2a and 2b). This was borne out by vertical profile measurements of tracer concentrations over the Øresund during one experiment (Gryning, 1985). The measurements revealed that the maximum concentrations occurred at a height of 50 m which is about 50 m below the release height. Thus, most of the tracer was situated immediately above the stable surface layer. The stable layer over the water undoubtedly tended to minimize mixing of the tracer into it. Finally, the fraction that became entrained within the stable layer was subjected to upward vertical motion as the surface air decelerated near the downwind coastline. The combined effects of these processes would tend to reduce the surface air concentration measured along the downwind coastline relative to those expected in the absence of the stable layer. Since the Matthew/Adpic models have no capability of including the effects of this stable layer, the calculated concentrations were expected to be higher than those measured. The reason for the close agreement between the calculated and measured concentration for the 5 June experiment is not at all clear. Vertical profiles of temperature over the water certainly detected the presence of the stable surface layer over the Øresund.

Thykier-Nielsen, Gryning and Mikkelsen (1987) presented simulations from six of the Øresund tracer experiment with RIMPUFF (the Risø mesoscale puff diffusion model). Simulating the tracer experiment, it was not straightforward how the complicated structure of the atmospheric stability over Øresund should be presented to the model. It was chosen to let the dispersion over the whole the Øresund area be controlled by the conditions at Barsebäck. Inland from the coast of Sjælland, dispersion characteristics were controlled by the conditions at Gladsaxe. The procedure did not take into account the stable layer close to the water surface over Øresund that inhibited dispersion and affected the wind field in the lower part of the atmosphere. However, above this stable layer the air was neutral or unstable and corresponded to upwind land conditions, although the turbulence was weaker because of decaying.

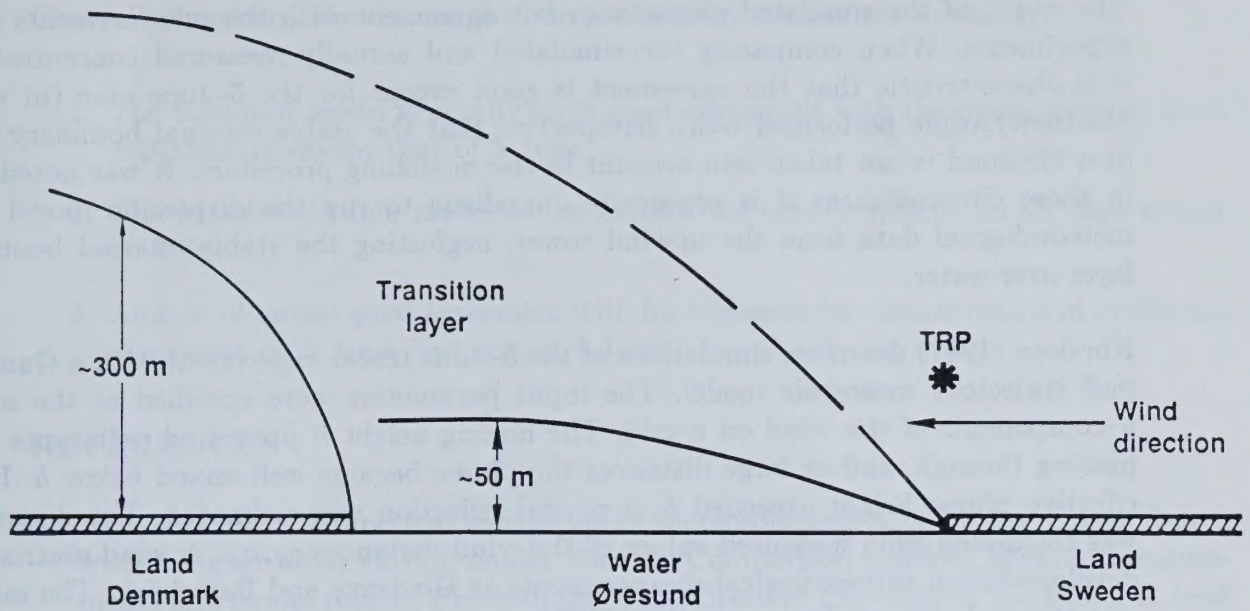


Figure 2.a: Generalized structure of the atmospheric boundary layers along an east-west cross section of the Øresund strait.

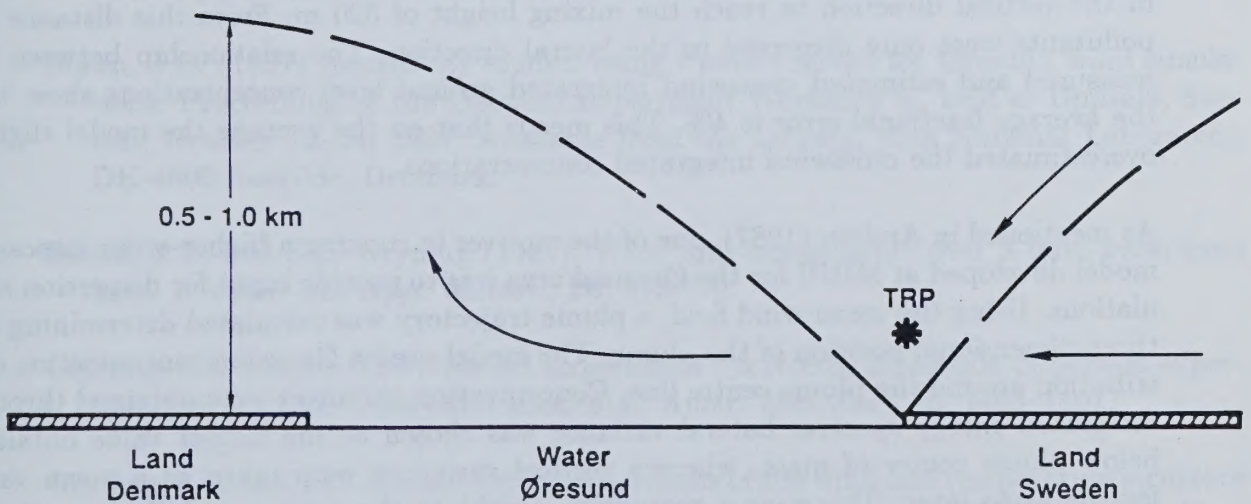


Figure 2.b: Generalized depiction of the atmospheric boundary layer flows along an east-west cross section of the Øresund strait.

In all the experiments there is good agreement between the position of the measured and simulated plumes. This shows that the wind field was well predicted by the model. The width of the simulated plume is in fair agreement with the measurements in all experiments. When comparing the simulated and actually measured concentrations, it is characteristic that the agreement is good except for the 5-June case (in which Matthew/Adpic performed well), irrespective that the stable internal boundary layer over Øresund is not taken into account in the modelling procedure. It was noted that in these circumstances it is physically unrealistic to run the dispersion model with meteorological data from the upwind tower, neglecting the stable internal boundary layer over water.

Knudsen (1987) describes simulations of the 5-June tracer experiment with a Gaussian puff trajectory mesoscale model. The input parameters were specified as the u and v -components of the wind on a grid. The mixing height h prevented pollutants from passing through, and at large distances the plume became well mixed below h . If the effective plume height exceeded h , a partial reflection was estimated. The dispersion was calculated from measured values of the wind variances $\sigma_\theta, \sigma_\phi$. A wind matrix was developed from meteorological measurements at Gladsaxe and Barsebäck. The matrix was oriented along a line connecting Gladsaxe and Barsebäck with a grid spacing of 5 km. The total area was 30×65 square kilometres. The σ -values were taken from the 10-min measurements at the Gladsaxe tower.

Two measured concentration traverses were compared with the modelling results, both in the city of Copenhagen. One traverse was measured by NILU (Sivertsen, 1986), and one by SCK/CEN (Vanderborght, 1986). The mixing height was estimated from the depth of the internal boundary layer at the distance where the concentration was to be estimated, i.e. $h = 300$ m for the NILU and $h = 200$ m for the SCN/CEN measurements.

Calculations show that at a downwind distance of about 6 km the plume had spread in the vertical direction to reach the mixing height of 300 m. From this distance the pollutants were only dispersed in the lateral direction. The relationship between the measured and estimated crosswind integrated ground level concentrations show that the average fractional error is 4%. This means that on the average the model slightly overestimated the crosswind integrated concentrations.

As mentioned in Andrén (1987), one of the motives in running a higher-order mesoscale model developed at MIUU for the Øresund area was to provide input for dispersion simulations. Using the mean wind field, a plume trajectory was calculated determining the three-dimensional position of the plume. The model used a Gaussian concentration distribution around the plume centre line. Concentration variances were obtained through integrated energy spectra. Lateral variance was chosen as the largest value obtained below plume center of mass, whereas vertical variances were taken as a mean value for the same layer. This gave a reasonable weight to the various stabilities vertically encountered by the plume and also took account of the maximum width of the plume as it was mixed down to the surface.

On 5 June the tracer experiment was modelled using both the large and small model domain. Both simulations gave predicted concentrations in good agreement with those measured. The tracer dispersion experiment on 4 June was modelled with the small model domain only. The predicted concentrations were considerably lower than those

measured.

Generally speaking, the results of these preliminary studies are confusing.

1. The Risø puff model generally gives good agreement with the measurements in all experiments except that of 5 June.
2. The Matthew/Adpic model overestimates the concentrations on all days except on 5 June.
3. Andrén obtained good agreement with his higher-order closure model in predicting dispersion on 5 June, but not on 4 June.

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Exerpt of letter from Chris Doran, Pacific Northwest Laboratory,
Richland, WA 99352

Although I won't be attending the October workshop, I haven't completely forgotten our Oresund work. In the past few months I have been working a bit on incorporating a turbulent kinetic (TKE) closure scheme in Pielke's model, which is looking less like Pielke's model as time goes by. I started with some code provided by Ray Arritt, who used to be one of Pielke's students, and made some changes in that. The scheme is essentially one introduced by Mellor and Yamada a number of years ago. Even though it's not new, it seems to have some definite advantages over Pielke's older closure scheme, especially over complex terrain or in regions where there are sharp changes in surface heat fluxes. As a test for the revised code I ran the Oresund simulation again, using the same initial conditions used for our paper. Some results are enclosed.

The wind speed behavior looks similar to the earlier results with two exceptions. First, the extreme accelerations I found just west of Sweden are no longer present. These latest results seem a lot more reasonable than the earlier ones, and are consistent with my judgment that the old version of the model did not handle the transition from unstable to stable conditions as well as I wanted. The wind speeds are also a bit lower than before. I presume I could make them higher by adjusting the initial conditions, but there didn't seem to be much point in that.

The potential temperature profiles over land and water are generally similar to earlier results as well. The unstable layer over Denmark may be a bit deeper, but the differences are small.

I've also included a contour plot of some TKE values. When I plotted labels on the contours they were somewhat difficult to read, but starting with the closed loops over and west of the Oresund, the contour values are 500, 1500, 2500, 5000, and 10000 cm^2/s^2 as you move outward from the water area. The growth of the well-mixed turbulent layer over Denmark is evident, although one has to interpret this with some caution since the model is hydrostatic.

Finally, I plotted wind speeds at the first 6 grid levels of the model, 5, 10, 20, 40, 80, and 140 m. The decreasing effect of the inversion over water is easy to see here.

I don't see anything that's particularly surprising, but I thought you might be interested in some of the differences and extra information that's available from the TKE closure version of the model. It might also be worth comparing these results with what other models give.

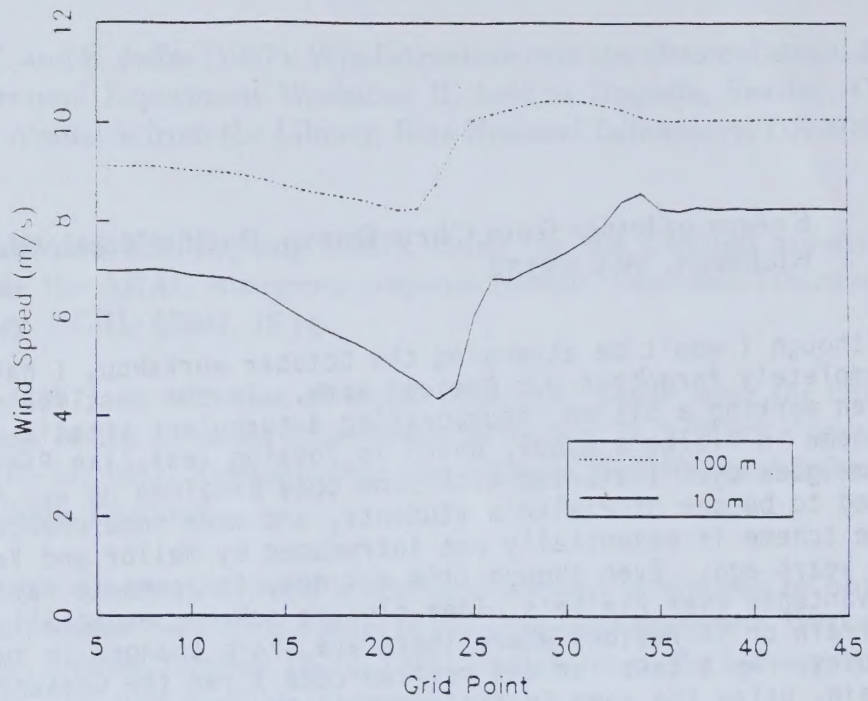


Figure 1. Simulated wind speeds at 10 and 100 m above the surface at 1200 CET, 5 June 1984. Oresund boundaries are grid points 25 and 34.

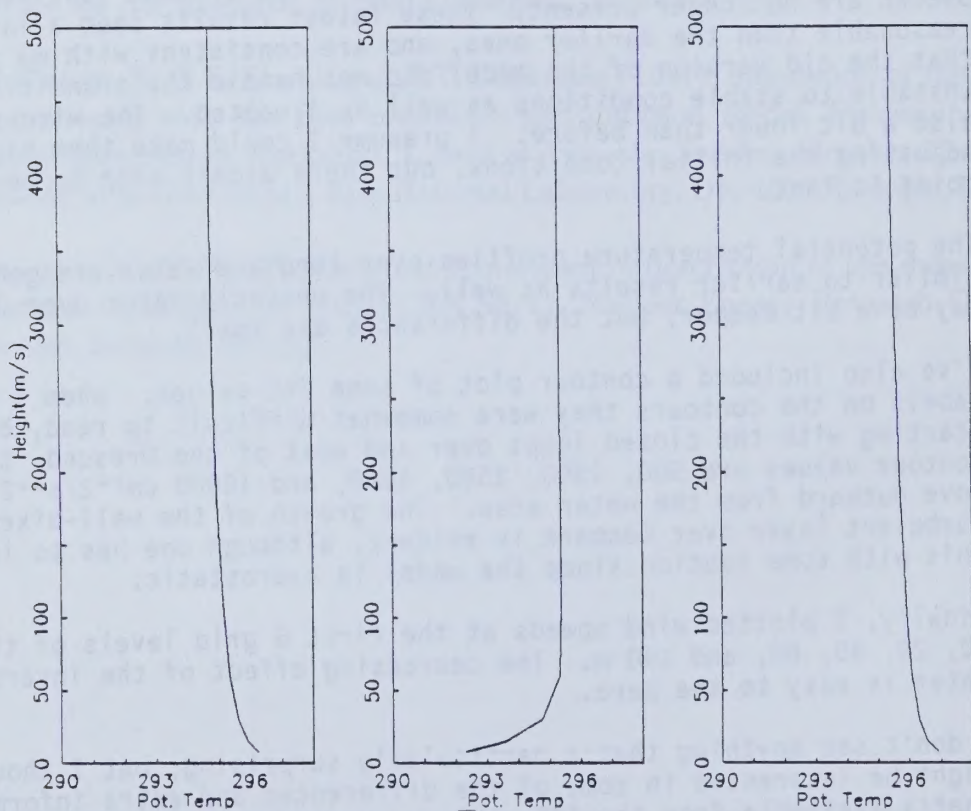


Figure 2. Simulated potential temperature profiles over Denmark (left), the middle of the Oresund (center), and Sweden (right).

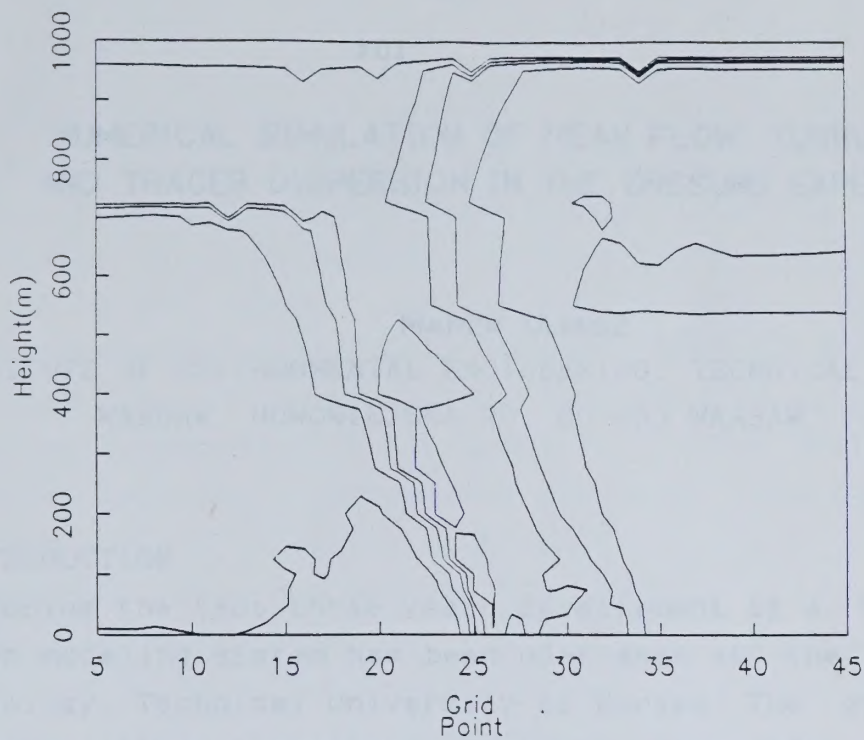


Figure 3. Simulated turbulent kinetic energy (TKE) contours. Starting with the closed loops over and to the west of the Oresund, the contour values are 500, 1500, 2500, 5000, and 1000 cm^2/s^2 .

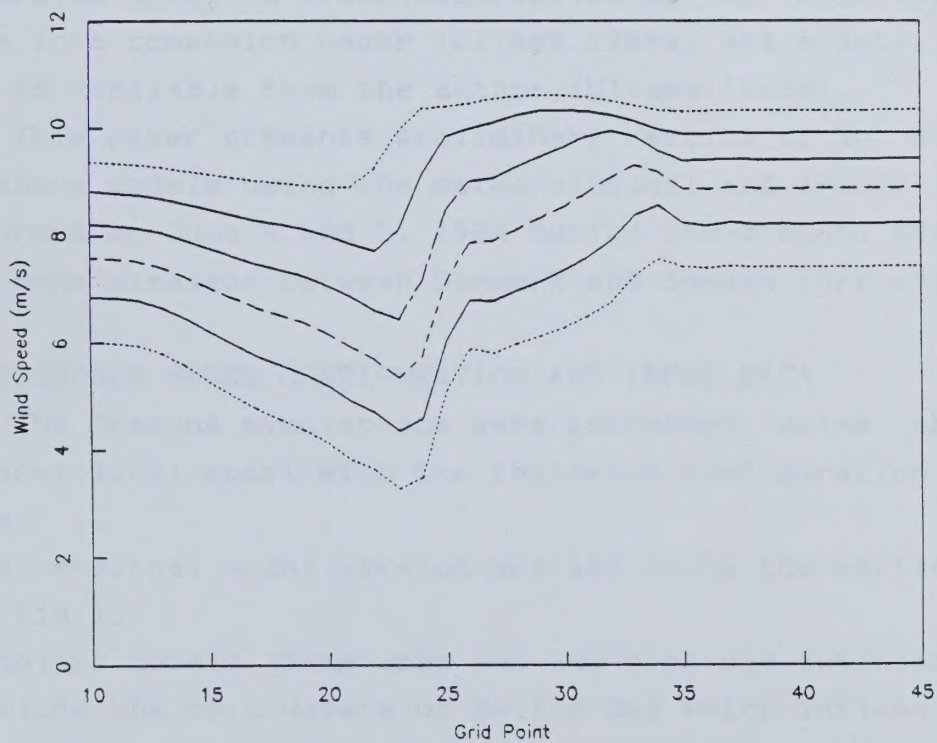
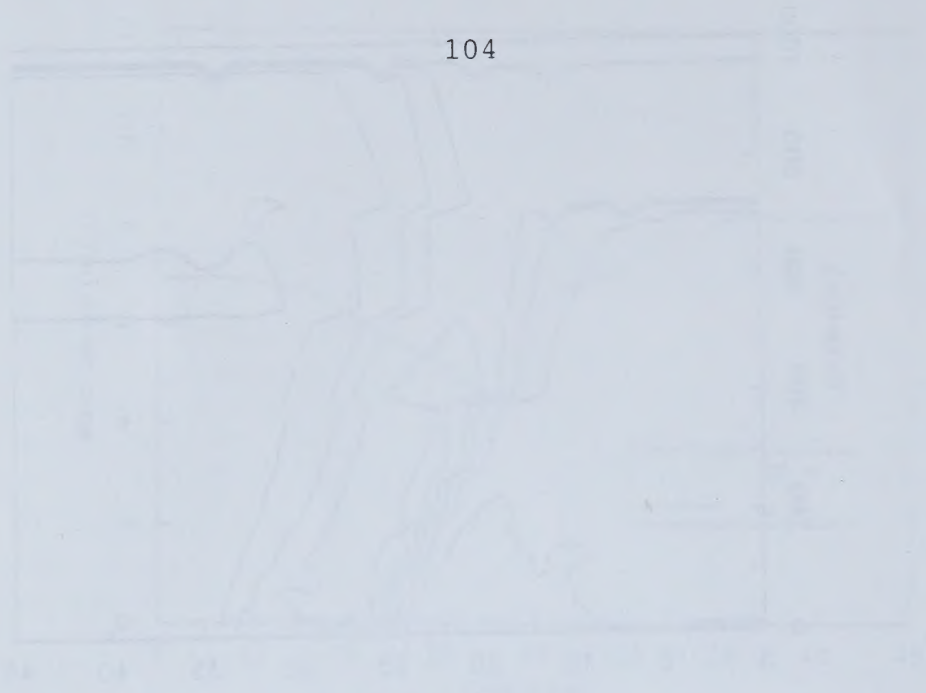


Figure 4. Simulated wind speeds at (top to bottom): 140 m, 80 m, 40 m, 20 m, 10 m, and 5 m.



to compare with the results of the other two methods. The results of the three methods are compared in Table I. The results of the three methods are compared in Table I. The results of the three methods are compared in Table I.



Figure 1. Comparison of the results of the three methods. The results of the three methods are compared in Table I. The results of the three methods are compared in Table I. The results of the three methods are compared in Table I.

NUMERICAL SIMULATION OF MEAN FLOW, TURBULENCE AND TRACER DISPERSION IN THE ØRESUND EXPERIMENT

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1. INTRODUCTION

During the last three years development of a mesoscale dispersion modeling system has been undertaken at the Department of Meteorology, Technical University of Warsaw. The system includes the 1-dimensional atmospheric boundary layer model, the 3-dimensional γ -mesoscale hydrostatic meteorological model with the second order turbulence closure and the Lagrangian particle/puff dispersion model. A brief description of the modeling system is given in a companion paper (Uliasz, 1988a) and a detailed description is available from the author (Uliasz, 1988b).

This paper presents preliminary results of an evaluation of the above models using the meteorological and tracer measurements performed on June 4 and 5, 1984 during the Øresund experiment over 20km wide straight between Denmark and Sweden (Gryning, 1985).

2. MESOSCALE MODEL CONFIGURATION AND INPUT DATA

The Øresund simulations were performed using the mesoscale meteorological model with the following configuration and assumptions:

- 2-dimensional model version applied along the section indicated in Fig.1;
- modeling domain 150km long and 2km high was extended eastward to include the cold waters of Baltic Sea which influences strongly the boundary layer structure over the Øresund (Anders, 1987);
- numerical mesh: 50*15 gridpoints with a horizontal step variable from 2km over the Øresund to 6km near lateral boundaries;
- integration time step $\Delta t = 120s$;
- variable surface roughness, but a flat terrain;

- simulations were performed from 6:00 to 13:00 CET using the upwind radiosounding at Borlunda to initialize the model.

The land surface skin temperature, T_s , was prescribed for the period of simulations as a function of local time t without spatial variability following Anders (1987) while sea surface temperature was kept constant. The geostrophic wind components, U_g and V_g , were assumed constant with height and time what is a serious simplification of synoptical conditions on June 4 and 5. These input data are summarized as follows:

	June 4	June 5
$T_s =$	$18 + 8 \sin(2\pi(t-6)/24)$	$22 + 7 \sin(2\pi(t-9.5)/24)$
$T_{sea} =$	12°C	12.5°C
$U_g =$	-16.0 m/s	-16.9 m/s
$V_g =$	4.4 m/s	5.9 m/s

Numerical simulations were carried out in three variants (called models A, B and D) using different turbulence parameterization in the mesoscale model and various configuration of the upwind part of the modeling domain:

simulation	turbulence closure	modeling domain
model A	level 2.5	} included Baltic Sea upwind
model B	level 2	
model D	level 2.5	infinite land upwind

The level 2.5 parameterization is based on the prognostic equation for turbulent kinetic energy while in the simpler level 2 parameterization turbulent energy is calculated diagnostically without taking into account advection processes. Since the mesoscale dispersion modeling system is developed using personal computers it is necessary to assess applicability of simpler and more computationally effective model formulations.

3. SIMULATION OF MEAN FLOW AND TURBULENCE

June 4 and 5 are characterized by moderate easterly winds over relatively cold water of the Öresund straight and warm land surfaces. The air is slightly unstable upwind from the Öresund over Swedish coast, then it gradually turns stable as it passes over the water surface. This stable stratification is quickly turned into an unstable one downwind from the Öresund over

Copenhagen.

Fig.2 presents simulated wind velocity and direction plotted against 1 hour averaged data from sodar and mast measurements. When passing over the Øresund, the wind at 100m gradually speeds up, reaching the maximum near the Danish coast then it decreases over the rough Copenhagen area. The surface wind has a different structure. The 10m wind decelerates over water reaching the minimum over Copenhagen. Over the more smooth farmland west of Copenhagen the wind speed again increases. In some cases (eg. June 4, 10:00) a small acceleration of surface wind ($\approx 0.5\text{m/s}$) over Swedish coast, when air meets a smooth water surface, is computed but it not so strong as reported in earlier Øresund data simulations by Doran and Gryning (1987). The simulated tendency in wind speed agree well with the measurement data. There are not significant differences in wind velocity predictions obtained from simulations A, B and D.

The surface wind direction is simulated with accuracy of 10-20 degrees. The formulation of the upwind conditions for the mesoscale model has a serious effect on the calculated wind directions as well as on the prediction of the mixing layer height (model A and D). The model D provides too high mixing layer as it is visible from the potential temperature profiles (Fig.3) over land and vertical profiles of turbulence (Fig.4) over water near Middelgrund. The reference simulation A seems to fit best aircraft measurements of turbulence with a characteristic minimum at height 100-200m.

The gradual lowering of turbulence as air passes the Øresund is well simulated by the model A with the level 2.5 turbulence parameterization (Fig.5). The model B with simpler turbulence closure predict too low turbulence intensity over the Øresund on June 4 while there is almost no difference between simulations A and B on June 5. The rapid growth of turbulence over Copenhagen is not so good reproduced. The model A responses too slowly for the surface forcing at least at heights 300-400m where aircraft measurements are available. The model B seems to be better in this situation.

4. SIMULATIONS OF TRACER DISPERSION

The tracer releases from 8:30 to 12:00 CET from the 100m Barseback tower on June 4 and 5 were simulated with the Lagrangian particle dispersion model. The model was run using the 2-dimensional fields of mean wind and velocity variances obtained in the meteorological simulations A and B described above.

Fig.6 presents configuration of sampling arcs and a tracer plume on 12:00, June 5 in XY and XZ sections simulated by releasing 4320 particles/hour. A low lateral dispersion of the plume over water and almost no mixing of the plume downward the water surface are well visible.

Tracer concentration at receptor points was calculated by counting particles in sampling volumes with sizes $250 \times 250 \times 50$ m during the period of 1 hour (11:00-12:00). Statistically stable results were obtained by averaging results of independent dispersion simulations performed with releasing 720 particles per hour. In order to compare predicted and measured concentrations the positions of receptor points were projected at the line perpendicular to the direction from the release point to the position of 1 hour maximum concentration measured at a given sampling arc.

The predicted values and sizes of concentration distributions agree relatively well with measurements on June 5 while the calculated concentrations on June 4 are 2-3 times too high (Fig.7). The model failure on June 4 tracer experiment can not be explained by incorrect prediction of wind and turbulence structure over the Öresund and must be related to oversimplifications of synoptical conditions in the mesoscale meteorological model. The model B which underestimates turbulence intensity over water results in higher concentration at sampling points than the model A. In all concentration predictions a certain shift of maximum concentration position appears due to inaccuracy in simulated wind direction.

CONCLUSIONS

The results of preliminary simulations indicate that the mesoscale model is able to correctly reproduce a general structure of the atmospheric boundary layer over the land-water-land area.

An improvement of the model results can be expected from (1) more realistic representation of synoptic conditions (geostrophic wind variable with time and height), (2) better representation of terrain (3-dimensional simulation including topography), and (3) including spatial variability of thermal forcing by prognostic calculation of the land surface temperature.

Acknowledgments

The author is grateful to Anders Andrén from Uppsala University, Sven-Erik Gryning and Niels Gylling Mortensen from Risø National Laboratory for providing data from the Öresund experiment.

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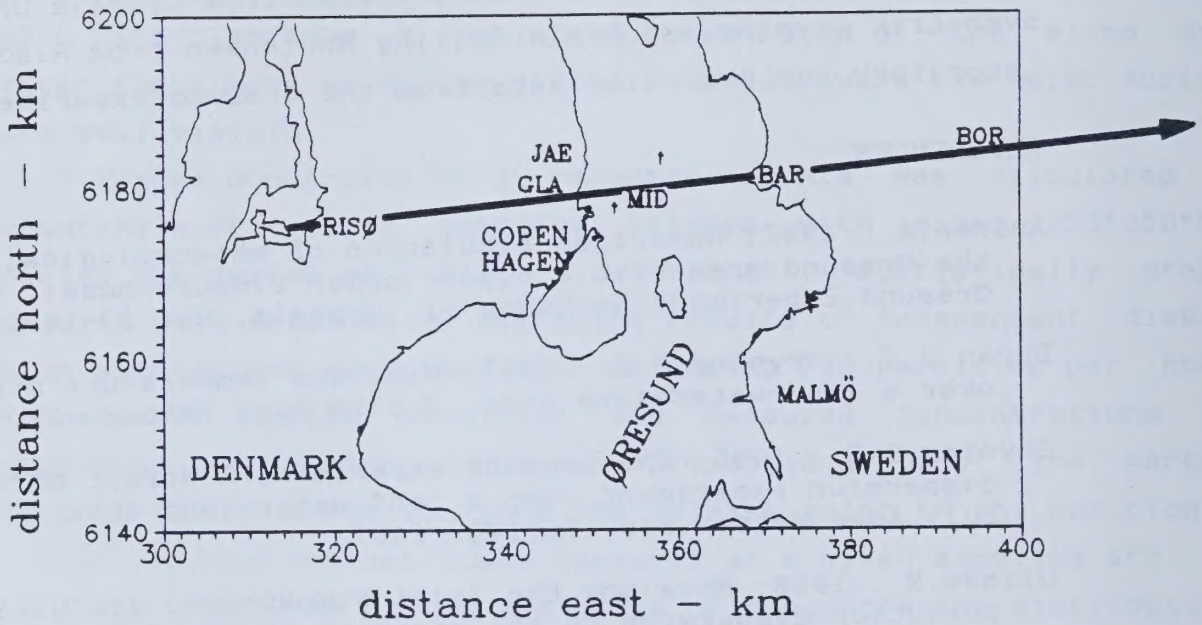


Fig.1

Map of the straight of Øresund and its surrounding (JAE - Jägersborg, GLA - Gladsaxe, MID - Middelgrund, BAR - Barsebäck, BOR - Borlunda, ↑ - minisonde releases from a fishing boat, modeling section indicated by an arrow)

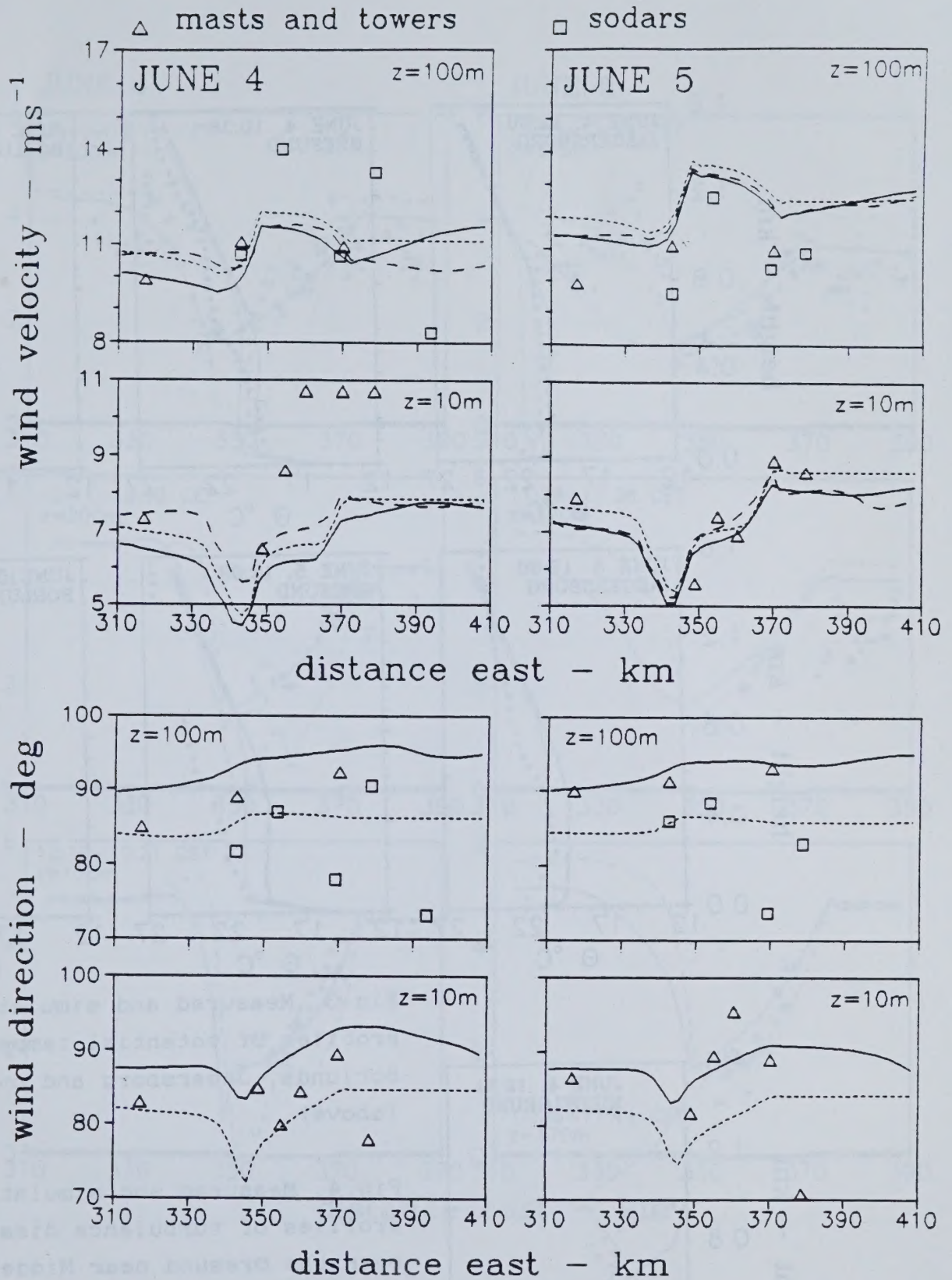


Fig.2. Horizontal profiles of wind speed and direction compared with sodar and mast measurements

model A - solid line

model B - long dashed line

model D - short dashed line

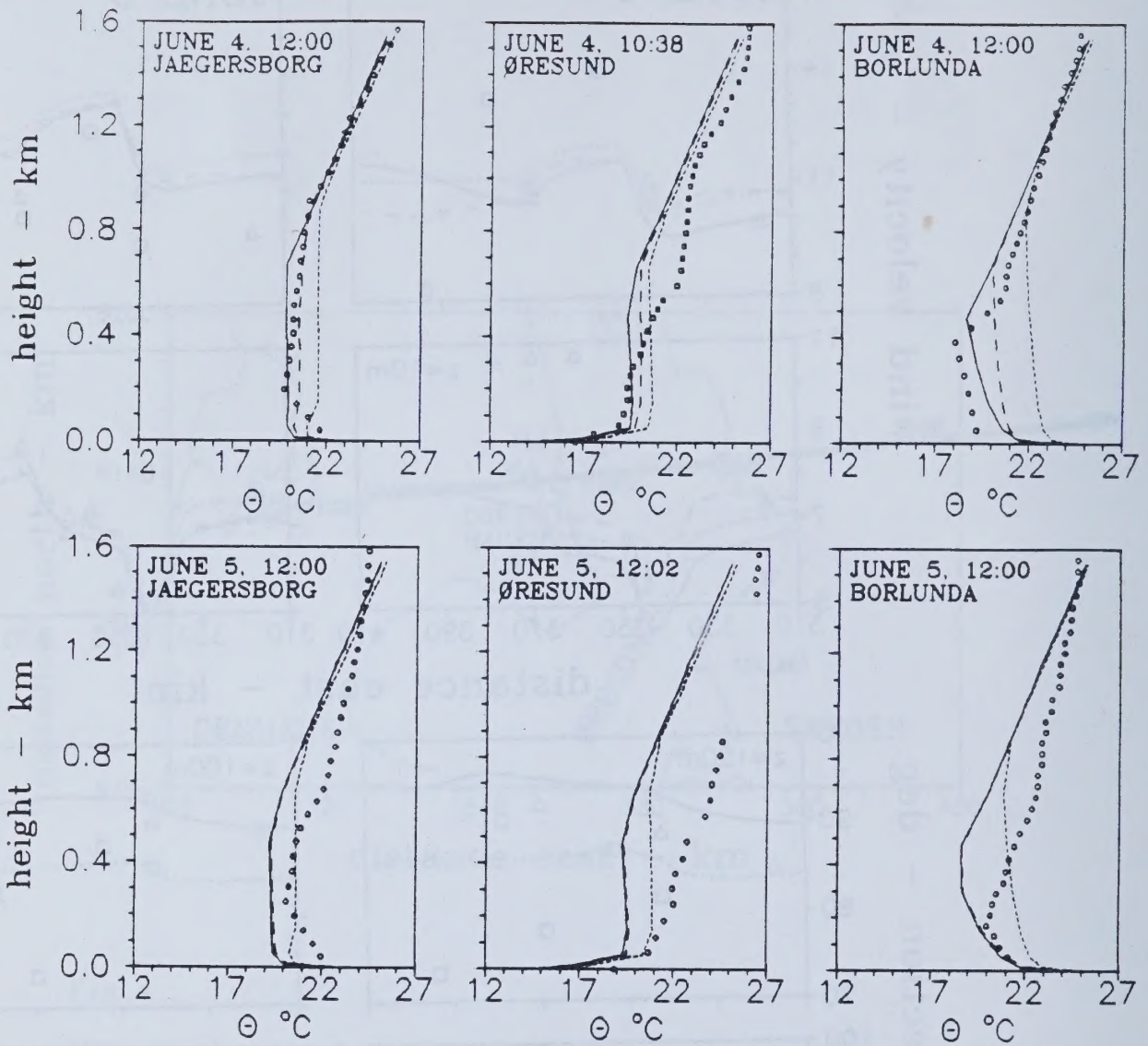


Fig.3. Measured and simulated vertical profiles of potential temperature at Borlunda, Jägersborg and over the Øresund (above)

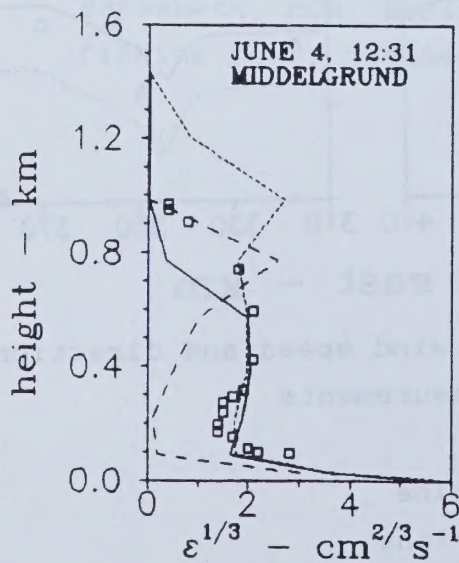


Fig.4. Measured and simulated vertical profiles of turbulence dissipation ratio over the Øresund near Middelgrund (left)

model A - solid line
 model B - long dashed line
 model D - short dashed line
 measurements - squares

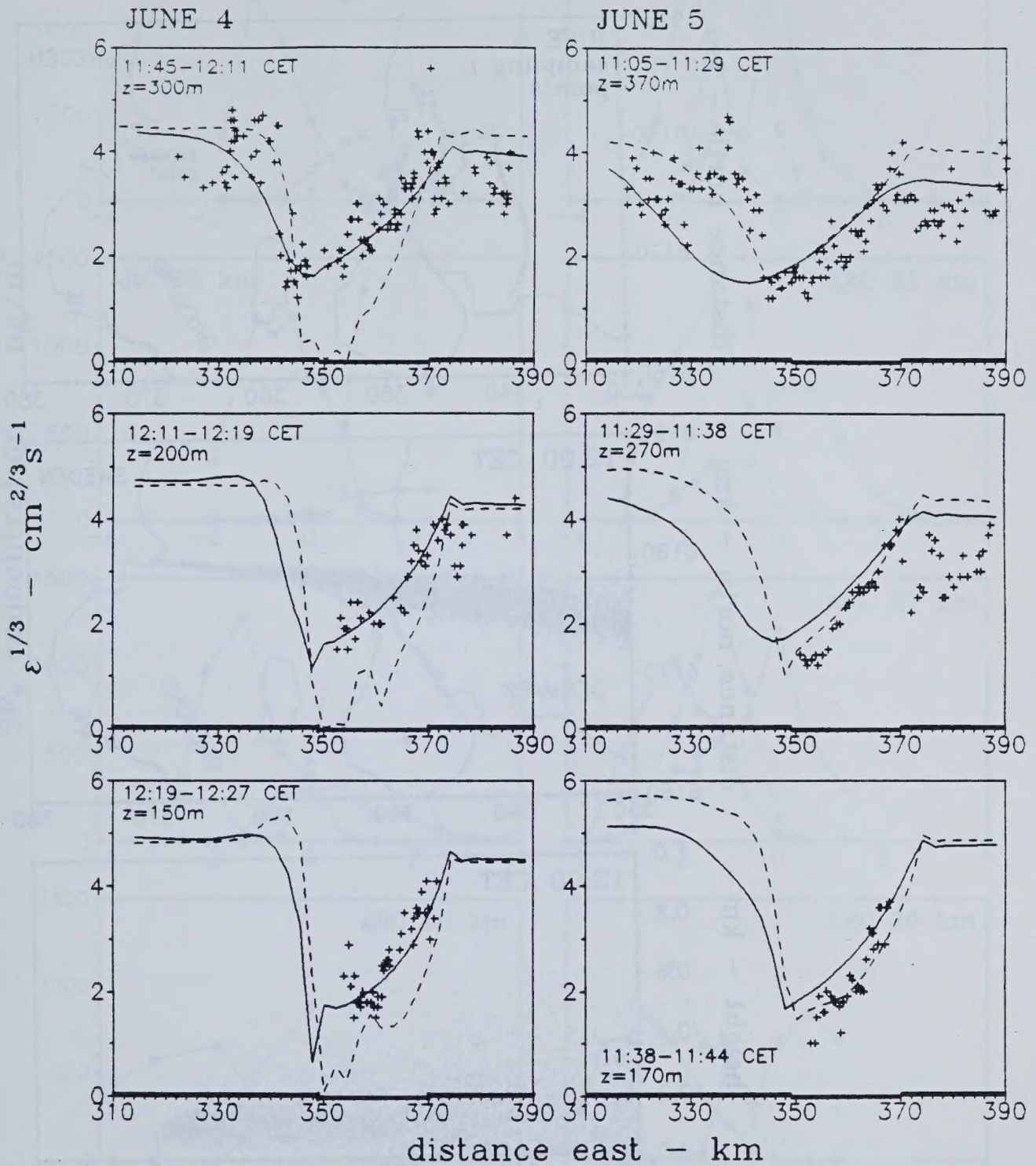


Fig.5. Measured and simulated turbulence dissipation ratio along west-east traverses over the Øresund (model A - solid line, model B - dashed line, aircraft measurements +)

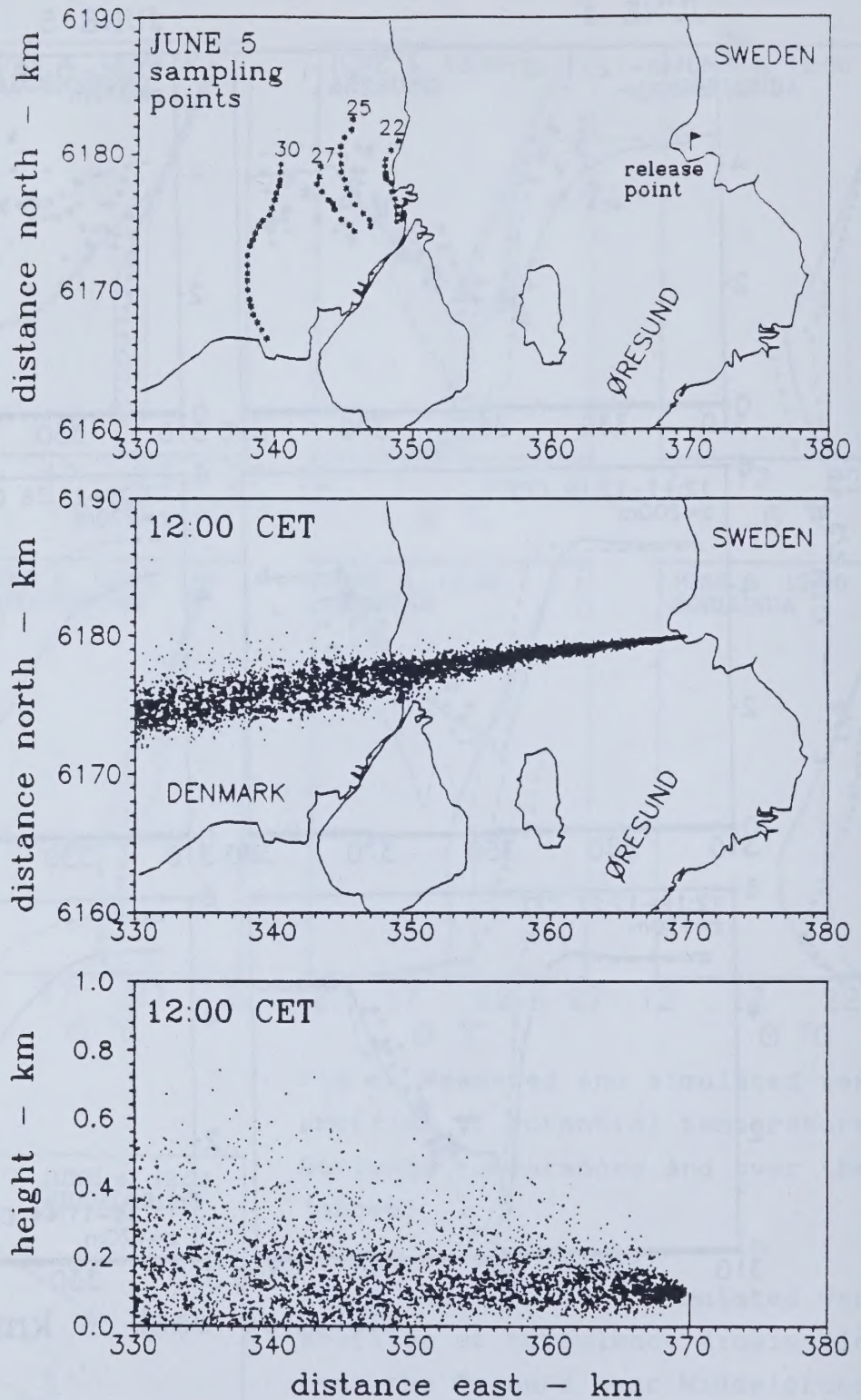


Fig.6

Sampling points configuration and tracer plume on 12:00, June 5 simulated by Lagrangian particle model

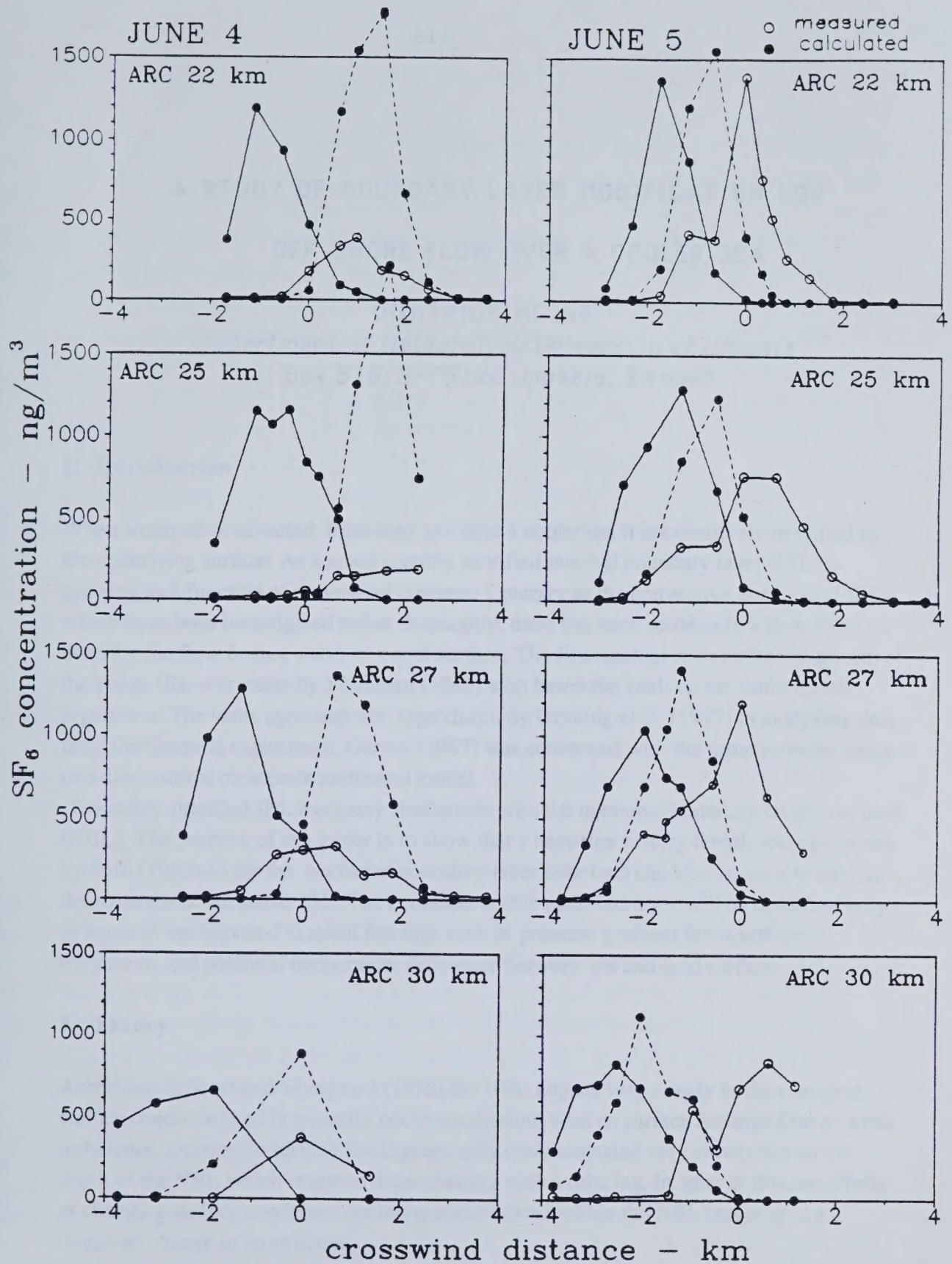


Fig.7. Measured and calculated tracer concentrations along sampling arcs indicated in Fig.6

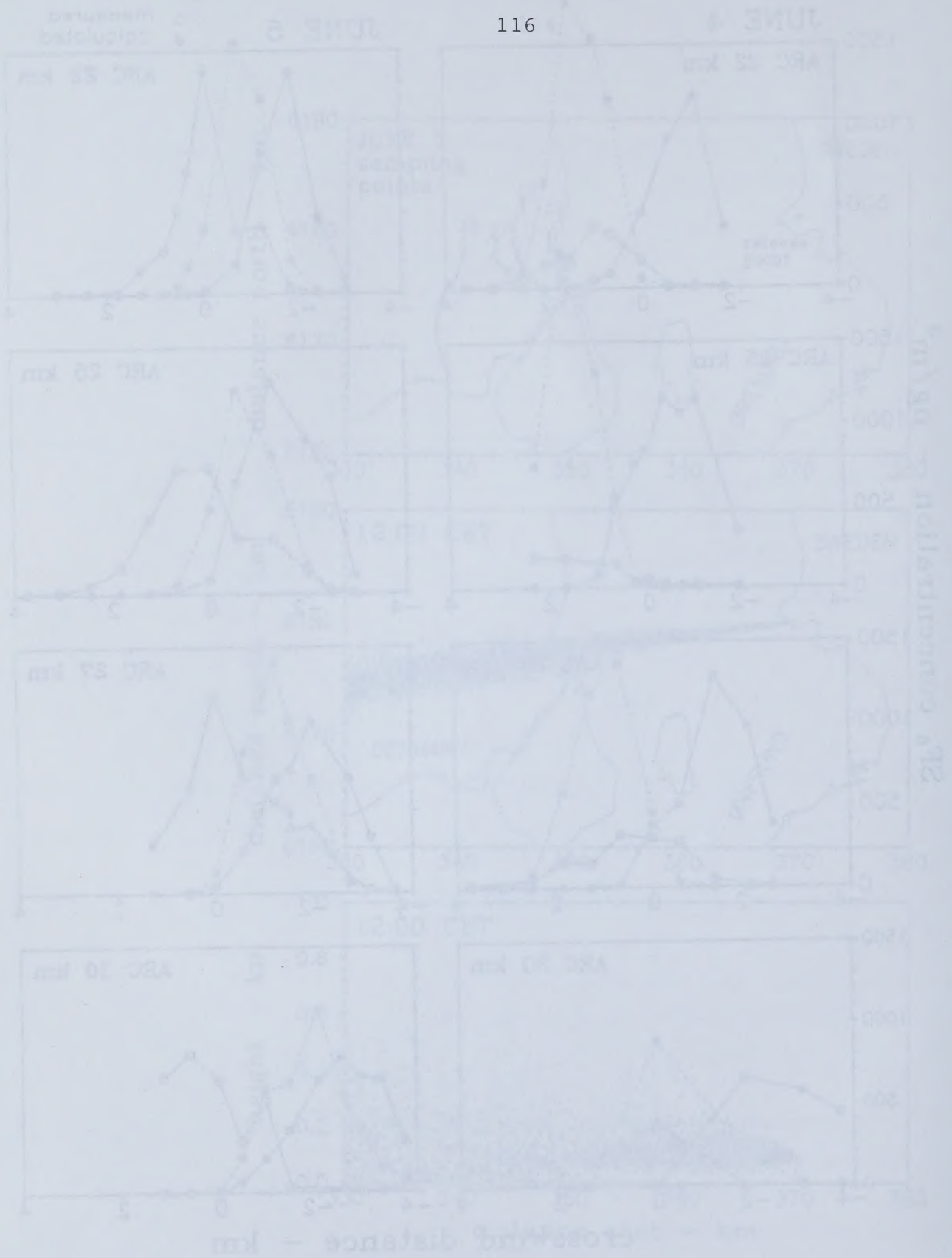


Fig. 2. Measured and calculated speed of movement along the axis of the river. The calculated speed is indicated in Fig. 2.

A STUDY OF BOUNDARY LAYER MODIFICATION FOR OFF-SHORE FLOW OVER A COOLER SEA

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1. Introduction

When warm air is advected from land out over a cooler sea it is extensively modified by the underlying surface. As a result a stably stratified internal boundary layer (IBL) is growing as a function of downwind distance. Contrary to the convective and neutral IBLs which have been investigated rather thoroughly, there has been made only a little effort to describe the flow from a warm to a cool surface. The first attempt to describe the growth of the stable IBL was made by Mulhearn (1981) who based the analysis on dimensional arguments. The same approach was reproduced by Gryning et al. (1987) in analyzing data from the Öresund experiment. Garratt (1987) was concerned with the same problem using a two dimensional mesoscale numerical model.

The stably stratified IBL has many similarities with the nocturnal boundary layer over land (NBL). The purpose of this paper is to show that a heat-flux history length scale proposed by Stull (1983a,b) for the nocturnal boundary layer over land can also be used to analyze the structure of the stable IBL. The evolution of this modified layer will be described only in terms of the imposed external forcings such as pressure gradient force, surface roughness, and potential temperature difference between sea and land surface.

2. Theory

According to Brost and Wyngaard (1978) the NBL adjusts very slowly to the changing surface conditions and is typically not in equilibrium with its surface forcings. Due to weak turbulence, changes in surface forcings are only communicated very slowly across the depth of the NBL which reacts to these changes with a time lag. In spite of this, the effects of changing surface conditions are being accumulated within the NBL resulting at a (lagging) change in its structure.

With such philosophy in mind Stull (1983a) suggested that the instantaneous heat flux may not be a governing parameter of the flow and perhaps an accumulation-type length scale based on the night-time heat-flux history is a relevant scale for the NBL temperature structure. This length scale is defined as

$$H = \int_0^t (Q_H / \rho C_p) d\tau / \Delta\theta_s \quad (1)$$

where t is the time since evening transition, Q_H is the surface heat flux, ρ is the density of air, C_p is the specific heat of air at constant pressure, τ is a dummy of integration, and $\Delta\theta_s$ is a temperature scale equal to the surface potential temperature change.

For the stably stratified IBL the definition of this length scale is modified to yield

$$H(x) = \int_0^x (Q_H(x) / \rho C_p) dx / (\bar{U}_r \Delta\theta_s) \quad (2)$$

where x is the distance downwind from the surface change, and $\bar{U}_r$ is a representative layer-average wind speed.

3. The data set

The theory proposed by Stull (1983a) has been tested against data from the Öresund experiment (Gryning 1985). One hour averaged data were used from the small mast on the Barsebäck shoreline, from the research vessel Aranda, and from the mast of Charlottenlund. The present analysis is restricted only to the 4th and 5th of June 1984 but in future work we plan to extend the analysis to all days with persistent easterly winds. The location of the measurement sites is shown in Figure 1.

During the intensive measuring period of the Öresund experiment the research vessel Aranda moved between five positions in the Öresund, making near surface measurements of wind and temperature fields. The prevailing winds during this period were persistently from the east and upstream (overland) profiles were measured at a small mast (12 m) erected at Barsebäck beach. Profiles of wind and temperature measured at the western coast of Öresund, at Charlottenlund site, are also used as representing overseas conditions. The friction velocity u_* and the temperature scale θ_* were computed for each site using a profile method and this allowed an interpolation of temperature to fixed levels. The roughness length of the water surface is estimated using Charnock's relation $z_0 = 0.0144 u_*^2 / g$.

Additionally the temperature profiles measured by minisonde ascents over the water, will be used to determine the height of the stable IBL.

Finally geostrophic winds for the 4th and 5th of June 1984 were calculated by Melas (1987) using an objective analysis of the pressure field in the Öresund region.

4. Parametrization of the length scale H

Gryning et al. (1987) showed that the behavior of the wind speed as air passes over the Öresund can be explained fairly well using a dimensionless distance $X = x(g\Delta\theta_s/\theta)/u^2$. In

the present study we have used $\Delta\theta_s = \theta_{Bb}(1.3) - \theta_w$ where θ_{Bb} is the potential temperature at Barsebäck mast and θ_w is the water temperature measured by Aranda, and $u = u_g$ where u_g is the component of the geostrophic wind normal to the coastal line. We expect that the properly nondimensionalised heat flux will be a function of the dimensionless distance. Stull (1983b) proposed a "turbulence generation" velocity scale $V_s = (f U_g z_0)^{1/2}$ which is related to the combined ability of pressure-gradient force (fU_g) and terrain roughness (z_0) to generate turbulence. $\Delta\theta_s$ may be a proper temperature scale.

The analysis of the pressure fields performed by Melas (1987) has shown that, during the 5th of June, calculated surface geostrophic winds are considerably lower than the winds measured by radiosonde ascents. This was attributed to baroclinicity and isalobaric winds. Thus it was decided to exclude these cases from our data set.

Figure 2 shows the dimensionless heat flux $u*\theta^*/(V_s \Delta\theta_s)$ as a function of the dimensionless distance X . A second order logarithmic curve has been fitted to the data

$$u*\theta^*/(V_s \Delta\theta_s) = 32.6 + 12.7 \ln(X+1) - 1.6 \ln^2(X+1) \quad (3)$$

The accumulated turbulent cooling within the stable IBL can now be determined by integrating the above equation over distance.

5. The potential temperature profile

Stull (1983a) proposed that the dimensionless potential temperature change, $\Delta\theta/\Delta\theta'_s$, is a function only of the dimensionless height, z/H . The potential temperature change at any height is defined as $\Delta\theta = \theta_{sea}(x,z) - \theta_{Bb}(z)$ and the surface potential temperature change as $\Delta\theta'_s = \theta_{sea}(1.3) - \theta_{Bb}(1.3)$. It was necessary to change the definition of the temperature scale used in the parametrization of H (eq. 3) so that $\Delta\theta/\Delta\theta'_s$ becomes equal to one when z/H becomes zero.

In order to isolate the advection effects on the temperature structure we neglect the temperature change with time. This is justified only during daytime conditions when the temperature difference between the land and the water surface is large.

Figure 3 shows the dimensionless potential temperature change plotted against the dimensionless height. An exponential curve provides a good eye-fit to the data.

$$\Delta\theta/\Delta\theta'_s = \exp(-0.75 z/H) \quad (4)$$

The coefficient $a=0.75$ in eq. 4 is surprisingly near the coefficient $a=0.77$ found by Stull (1983a) using Wangara and Koorin boundary layer data.

The exponential profile shape implies a negative curvature for the temperature profiles. Andre and Mahrt (1982) discussed the relative contributions of radiative and turbulent cooling to the development of the nocturnal surface inversion and suggested that strong

negative profile curvature is associated with a dominance of the radiative cooling.

Neglecting subsidence and the time change of temperature the potential temperature equation for the two dimensional case reads

$$\overline{U} \frac{\partial \overline{\theta}}{\partial x} = -\overline{\partial w \theta / \partial z} + \overline{S \theta} \quad (5)$$

where $\overline{S \theta}$ is the potential temperature change due to the net radiative flux divergence. Eq. 5 can be integrated over distance (from the shoreline) and over height (from surface to some fixed height well above the IBL height) to yield

$$H \Delta \theta_s \overline{U_r} / 0.75 = H \Delta \theta_s \overline{U_r} + \overline{R} \quad (6)$$

where $\overline{U}$ in eq.5 is replaced by $\overline{U_r}$, the representative layer-average wind speed, and $\overline{R}$ is the net radiative cooling for the whole IBL integrated over distance from the shoreline (clear-air radiative cooling above the stable IBL has been neglected). From eq. 2 we realize

that $H \Delta \theta_s \overline{U_r}$ is the accumulated cooling of the IBL due to turbulence transport and we may estimate that 25 % of the total cooling inside the stable IBL is due to clear-air radiational cooling and 75 % is due to turbulent transport. This indicates that the negative curvature of the temperature profiles in this case is not related to the relative magnitude of radiative cooling but is rather due to advection effects.

6. Depth of the stable IBL

The height of the top of the IBL, h , is defined as that height where the potential temperature change is 2 % of the surface value. Then the exponential profile implies that $h = 5.2 H$. Unfortunately, there are only two minisonde ascents over the Öresund which can be used to estimate h .

Figure 4 shows one of the minisonde profiles. The temperature profile reveals a strong inversion layer near water surface which is replaced by a weak elevated inversion above 70 m. This elevated inversion is due to large scale subsidence which does not affect our results considerably. The parameterized IBL depth, $h = 64$ m, provides a reasonable estimation for the depth of the surface based inversion layer. It is noticed that the minisonde was launched at a position about 12 km from the Swedish coast. This indicates that the stable IBL grows very slowly with an average fetch to height ratio of about 170. This must be compared with the results of Raynor et al. (1979) who found, for the unstable IBL over land, an average fetch to height ratio of about 30.

The same good agreement is achieved even for the second radiosonde profile.

Mulhearn (1981) suggested an empirical relationship for the height of the stable IBL

$$h = 0.0146 \times (g \Delta\theta_s / \theta u^2)^{-0.47} \quad (7)$$

Estimates of h made by eq. 7 are considerably lower than the values we get in the present analysis.

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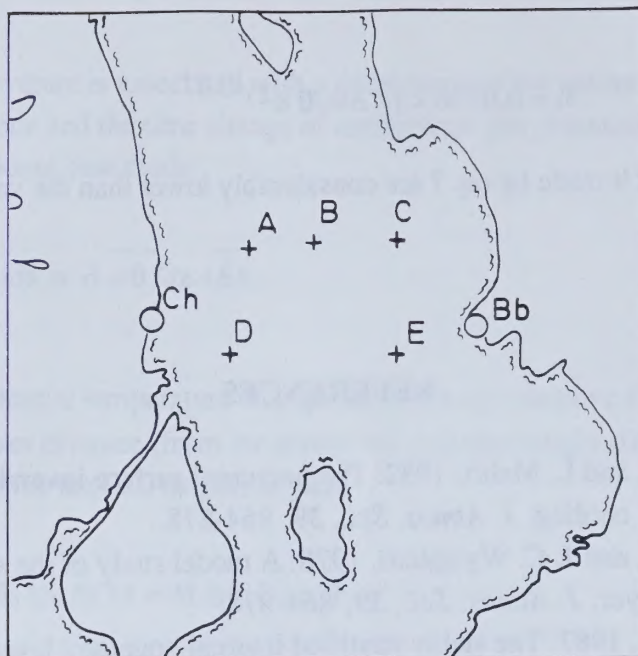


Figure 1. Map of the Öresund area, showing the measurement sites: Bb and Ch are the locations of the Barsebäck and Charlottenlund masts respectively, and A, B, C, D and E are the five positions where the research vessel Aranda performed measurements.

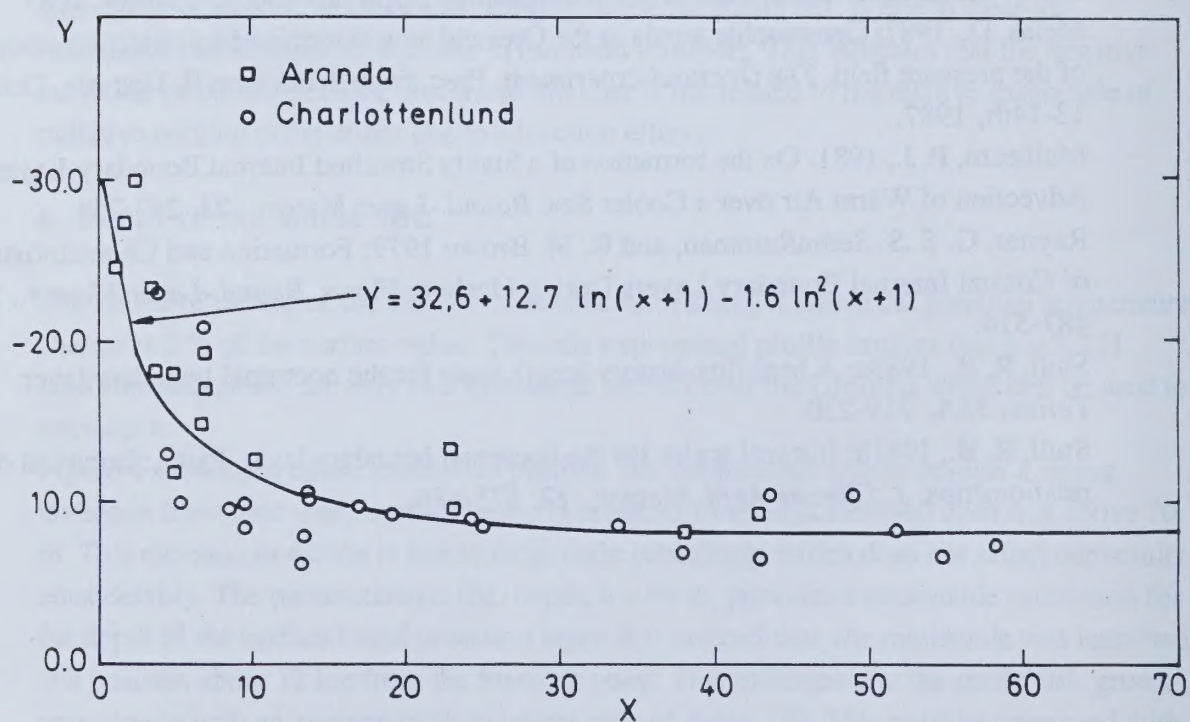


Figure 2. The dimensionless heat flux, $Y = u_* \theta_* / (V_s \Delta \theta_s)$, plotted as a function of the dimensionless water fetch $X = g(x \Delta \theta_s / \theta) / u_g^2$.

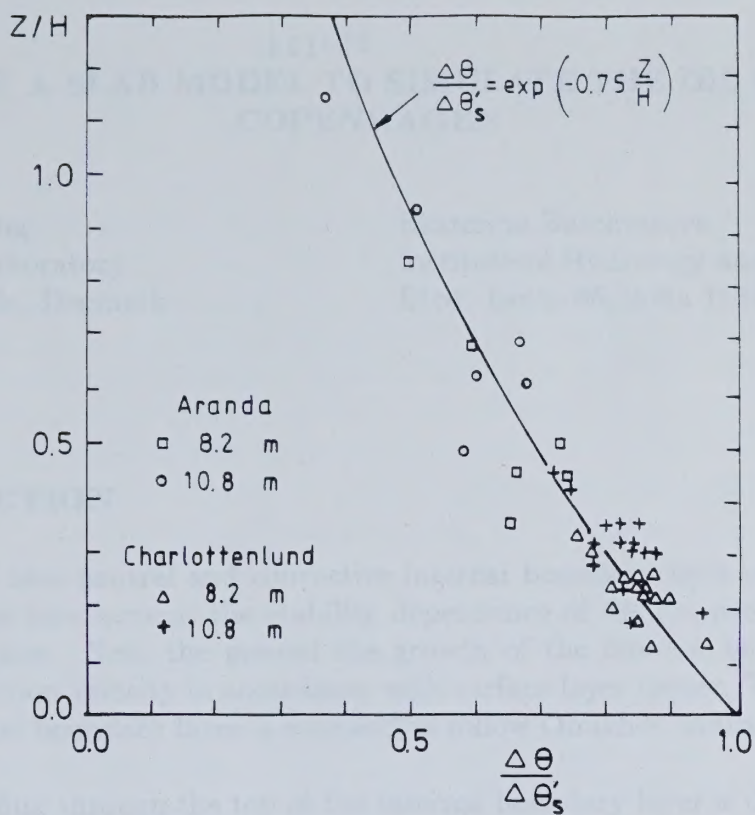


Figure 3. The dimensionless potential temperature change, $\Delta\theta/\Delta\theta'_s$, plotted against dimensionless height, z/H .

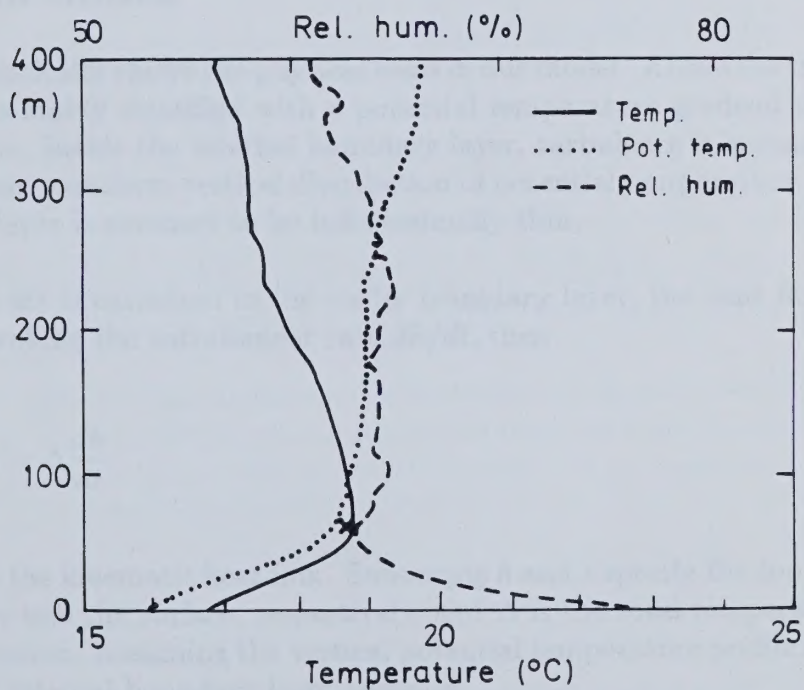


Figure 4. Minisonde profiles measured at 10.38 CET on 4th of June. The minisonde was launched from a fishing boat at about 12 km from the Swedish coast. The cross shows the parameterized depth of the stable IBL.

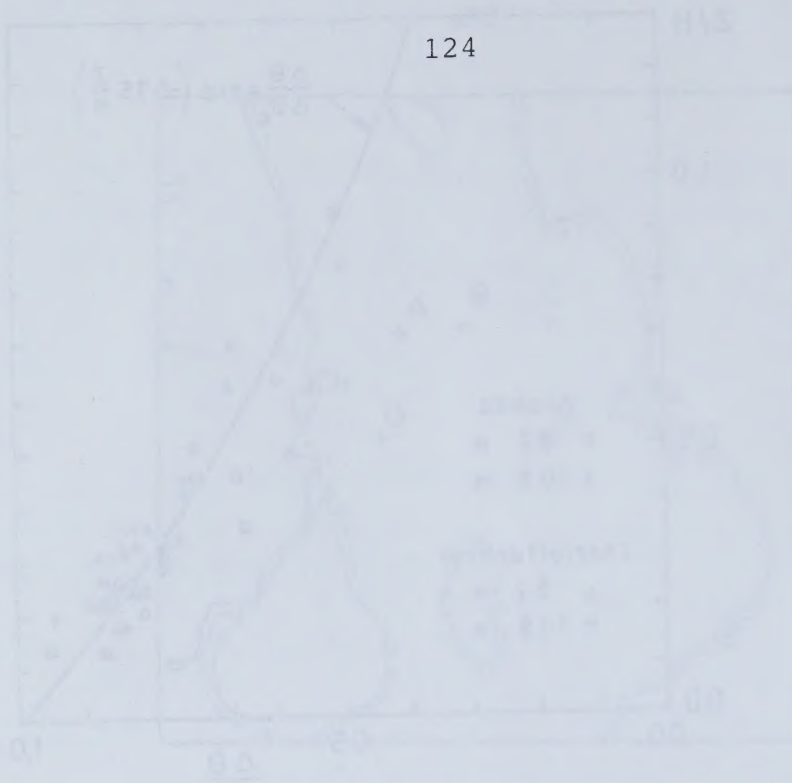


Figure 1. Map of the study area showing the distribution of the species. The map includes a grid with latitude (0° to 10° N) and longitude (10° to 20° E) coordinates. A legend indicates the following data series: 'Total' (solid line), 'A' (dashed line), 'B' (dotted line), 'C' (dash-dot line), 'D' (long-dashed line), and 'E' (short-dashed line). The map shows various contour lines and data points across the region.

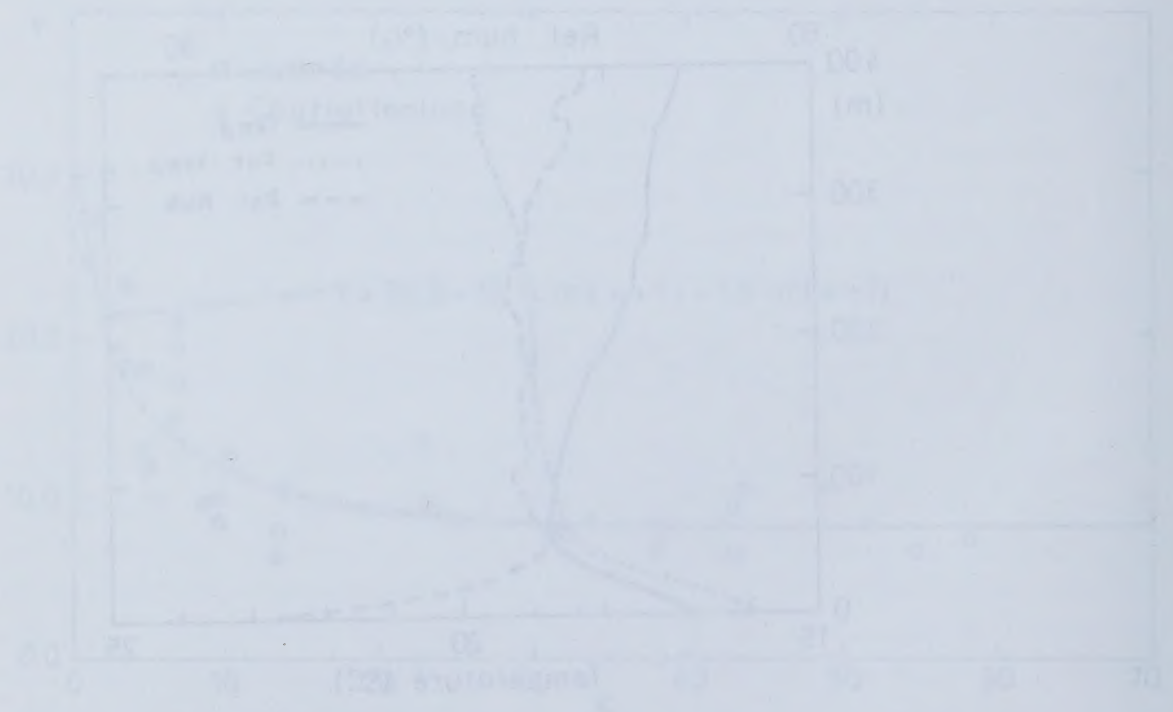


Figure 2. Graph showing the relationship between Temperature (°C) on the x-axis (0 to 30) and Relative Humidity (%) on the y-axis (0 to 100). The graph includes several curves and data points. A legend indicates the following data series: 'Total' (solid line), 'A' (dashed line), 'B' (dotted line), 'C' (dash-dot line), 'D' (long-dashed line), and 'E' (short-dashed line). The graph shows how relative humidity changes with temperature for different species or conditions.

USE OF A SLAB MODEL TO SIMULATE THE IBL OVER COPENHAGEN

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1 INTRODUCTION

A slab model of near-neutral and convective internal boundary layer evolution is presented that takes into account the stability dependence of the temperature difference across the inversion. Near the ground the growth of the internal boundary layer is controlled by friction velocity in accordance with surface layer theory. The wind profile inside the internal boundary layer is assumed to follow Obukhov similarity theory.

Kinematic heat flux through the top of the internal boundary layer is described by the formulation $-(\overline{w'\theta'})_h = 0.2(\overline{w'\theta'})_s + 2.5 u_*^3 T / gh + \text{Zilitinkevich correction}$. Simulated observations of near-neutral internal boundary layer heights carried out as part of the Øresund-experiment yielded good agreement between measurements and predictions.

2 THE SLAB MODEL

Figure 1 schematically shows the physical basis of our model. Above the internal boundary layer, air is stably stratified with a potential temperature gradient independent of height and time. Inside the internal boundary layer, turbulence is assumed to be sufficient to maintain a uniform vertical distribution of potential temperature. The inversion that caps the layer is assumed to be infinitesimally thin.

Because warm air is entrained in the cooler boundary layer, the heat flux at its top is downward. Invoking the entrainment rate dh/dt , then

$$-(\overline{w'\theta'})_h = \Delta \frac{dh}{dt}, \quad (1)$$

where $(\overline{w'\theta'})$ is the kinematic heat flux. Subscripts h and s specify the top of the internal boundary layer and the surface, respectively, and Δ is the total temperature difference across the inversion. Assuming the vertical potential temperature profile is uniform, the heating of the internal boundary layer becomes

$$\frac{d\theta}{dt} = \frac{(\overline{w'\theta'})_s}{h} - \frac{(\overline{w'\theta'})_h}{h}, \quad (2)$$

where θ is potential temperature and t time. As the heat flux at the inversion is negative it contributes to the heating of the layer, the formulation

$$-(\overline{w'\theta'})_h = A(\overline{w'\theta'})_s + \frac{Bu_*^3 T}{gh} - \frac{Bu_*^2 T}{Cgh} \frac{dh}{dt} \quad (3)$$

is applicable. Here, u_* is the friction velocity, g/T the buoyancy parameter and A, B , and C constants. This is derived from the turbulent energy budget in the entrainment zone above a boundary layer over a horizontally homogeneous area (Driedonks and Tennekes, 1981). The terms $A(\overline{w'\theta'})_s$ and $Bu_*^3 T/gh$ account for convective and mechanical turbulence, respectively. When h is small, the internal boundary layer rapidly entrains. To maintain a finite growth rate, the time derivative of the turbulent kinetic energy is included, called the Zilitinkevich correction (Zilitinkevich, 1975). It follows from Eq. (3) that close to the ground

$$\frac{dh}{dt} = Cu_* \quad (4)$$

When $C = 1.3$, Eq.(4) conforms to Miyake's theory for the growth of an internal boundary layer in the neutral surface layer (after Panofsky and Dutton, 1984). By introducing the Obukhov length

$$L = -\frac{u_*^3 T}{\kappa g(\overline{w'\theta'})_s} \quad (5)$$

where κ is the von Karman constant, Eq.(3) can be written:

$$(\overline{w'\theta'})_s - (\overline{w'\theta'})_h = (1 + A)(\overline{w'\theta'})_s \left(h - \frac{B\kappa L}{1 + A} \right) h^{-1} - \frac{Bu_*^2 T}{Cgh} \frac{dh}{dt} \quad (6)$$

Tennekes (1973) estimated A to be 0.2 and $B = 2.5$ was suggested by Tennekes (1973), van Dop et al. (1982), and Gryning and Batchvarova (1988).

The inversion strength, Δ , will tend to decrease as the boundary layer is heated due to entrainment of warm air from above and heating of the ground. It will tend to increase due to entrainment of the internal boundary into stable air above. Therefore

$$\frac{d\Delta}{dt} = \gamma \frac{dh}{dt} - \frac{d\theta}{dt} \quad (7)$$

where γ is the potential temperature gradient above the layer. These are the governing equations.

3 INVERSION STRENGTH

An expression of the temperature difference across the inversion is derived by substituting Eqs. (1) and (2) into Eq. (7):

$$\left(h \frac{d\Delta}{dh} + \Delta - \gamma h\right) \frac{dh}{dt} = -(\overline{w'\theta'})_s \quad (8)$$

and then from Eq. (1):

$$h \frac{d\Delta}{dh} + \Delta - \gamma h = -\Delta \frac{(\overline{w'\theta'})_s}{A(\overline{w'\theta'})_s + \frac{Bu_*^3 T}{gh}} \quad (9)$$

In Eq. (9) the Zilitinkevich correction, which is important only near the surface, has been omitted. By introducing the Obukhov length, Eq. (9) can be rewritten as

$$\frac{d\Delta}{dh} + \Delta \left(\frac{1}{h} + \frac{1}{Ah - B\kappa L} \right) = \gamma \quad (10)$$

This is an ordinary linear first-order differential equation; the solution is given in Gryning and Batchvarova (1988).

Figure 2 shows the exact analytical solution for dimensionless inversion strength, $\frac{\Delta}{\gamma h}$, plotted as a function of the stability parameter $\frac{h}{L}$ for $A = 0.2$ and $B = 2.5$. A simple approximation with correct asymptotic limits for neutral and convective conditions

$$\frac{\Delta}{\gamma h} = \frac{Ah - B\kappa L}{(1 + 2A)h - 2B\kappa L} \quad (11)$$

is also shown. The equation for the temperature difference across the inversion allow derivation of the heating rate of the internal layer. Writing Eq. (7) as:

$$\frac{d\theta}{dh} = \gamma - \frac{d\Delta}{dh} \quad (12)$$

and using Eq. (10), it can be transformed to:

$$\frac{d\theta}{dh} = \frac{\Delta}{h} \left(\frac{(1 + A)h - B\kappa L}{Ah - B\kappa L} \right) \quad (13)$$

A rather simple expression results when the approximation Eq. (11) is substituted into Eq. (13):

$$\frac{d\theta}{dh} = \frac{(1 + A)h - B\kappa L}{(1 + 2A)h - 2B\kappa L} \gamma \quad (14)$$

which constitutes the relationship between heating and growth of the internal boundary layer as function of its height, the atmospheric stability inside and the potential temperature gradient outside the layer.

4 HEIGHT OF THE INTERNAL BOUNDARY LAYER

Derivation of a prediction equation for h as function of downwind distance from the coastline involves the basic equation of heat conservation

$$\frac{\partial(u\theta)}{\partial x} + \frac{\partial(w\theta)}{\partial z} = -\frac{\partial(\overline{w'\theta'})}{\partial z} \quad (15)$$

and the continuity equation

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \quad , \quad (16)$$

where u and w are the wind velocities in the x - and z -direction, respectively. A detailed description of the derivation is given in Gryning and Batchvarova (1988). Briefly, Eq.(15) is integrated from $z = 0$ to $h(x)$ taking into account the boundary conditions for w and θ at the top of the internal boundary layer. Integration of Eq.(16) with respect to z permits the boundary conditions for w ; the boundary conditions for θ is obtained from the assumption of constant potential temperature with height. The result is:

$$\frac{d\theta}{dx} \int_0^{h(x)} u \, dz = (\overline{w'\theta'})_s - (\overline{w'\theta'})_h \quad , \quad (17)$$

which can be written by use of Eq. (14):

$$\left\{ \gamma \left(\frac{(1+A)h - B\kappa L}{(1+2A)h - 2B\kappa L} \right) \int_0^{h(x)} u \, dz \right\} \frac{dh}{dx} = (\overline{w'\theta'})_s - (\overline{w'\theta'})_h \quad . \quad (18)$$

To calculate the integral in Eq. (18), the wind profile within the internal boundary layer must be known. Little information, however, is available on this profile. However, describing the wind velocity profile by Obukhov similarity theory (strictly, valid only under homogeneous, stationary conditions) yields:

$$\int_0^{h(x)} u \, dz = \left(u_h - R \left(\frac{h}{L} \right) u_* \right) h \quad (19)$$

where u_h is the wind velocity in the upper part of the internal boundary layer and

$$R\left(\frac{h}{L}\right) = \frac{1 - (1 - 16\frac{h}{L})^{3/4}}{12\kappa(\frac{h}{L})} \quad (20)$$

(Gryning and Batchvarova, 1988). Figure 3 shows R as function of the stability parameter h/L . Under very convective conditions $R = 0$, and in a completely neutral atmosphere $R = 1/\kappa$. Inserting Eqs. (6) and (19) into Eq. (18) leads to the differential equation for h

$$\left\{ \left(\frac{h^2}{(1+2A)h - 2B\kappa L} \right) \left(u_h - R\left(\frac{h}{L}\right) u_* \right) + \frac{Bu_*^2 u_{hT}}{\gamma Cg(1+A)h - B\kappa L} \right\} \frac{dh}{dx} = \frac{(w'\theta')_s}{\gamma} \quad (21)$$

By neglecting the rather weak dependence in the R -function of h and with the boundary condition $h = 0$ at $x = 0$, we obtain

$$\begin{aligned} \frac{(u_h - Ru_*)}{1+2A} \left\{ \frac{h^2}{2} + \frac{2B\kappa L}{1+2A} h + \left(\frac{2B\kappa L}{1+2A} \right)^2 \ln \left(-\frac{1+2A}{2B\kappa L} h + 1 \right) \right\} \\ + \frac{Bu_*^2 u_{hT}}{C\gamma g(1+A)} \ln \left(-\frac{1+A}{B\kappa L} h + 1 \right) = \frac{(\overline{w'\theta'})_s}{\gamma} x \quad (22) \end{aligned}$$

A characteristic value of R is determined from Eq. (20) by inserting the actual Obukhov length and a typical value of the internal boundary layer height.

5 COMPARISON WITH DATA

The model for the growth of the internal boundary layer is tested on data from the Øresund-experiment. The growth of the internal boundary layer over Copenhagen was simulated on 29 May and 4-5 June, 1984. On those days the wind blew over the water towards Copenhagen. The height of the internal boundary layer over Copenhagen was determined by Batchvarova and Gryning (1988), Table 1, from measurements carried out by airplane. Turbulence, measured at 115 m, 5 km inland over Copenhagen, provided

Table 1: Values of the height of the internal boundary layer and land fetch.

Experiment 1984	Time CET	Internal boundary layer height (m)	over-land fetch (m)
29 May	1140	310	6400
4 June	1150	270	6300
5 June	1110	310	4900

the friction velocity and wind speed. Heat flux was measured at 10 m over the Swedish coast east of Copenhagen, in an area somewhat similar to suburban Copenhagen. Radio soundings carried out in the middle of the Øresund Strait show that the air over the water surface is (strongly) stably stratified up to a height of 50-100 m (Batchvarova and Gryning, 1988). Above, the air is also stable but the temperature gradient is less. To take into account the height dependence of the temperature gradient over the water, the model simulations were carried out in two height intervals. Up to a height of 100 m the characteristic temperature gradient was used in the layer of very stable air near the water surface; above 100 m the temperature gradient in the upper layer was used. The parameters used in the simulations are given in Table 2.

Table 2: Input data for the model simulation.

Experiment	Time	u_h	u_*	$(\overline{w'\theta'})_0$	Γ_1	Γ_2
1984	CET	(ms^{-1})	(ms^{-1})	(mKs^{-1})	(Km^{-1})	(Km^{-1})
29 May	1140	7.55	0.78	0.058	0.039	0.0041
4 June	1150	10.7	0.84	0.107	0.040	0.0042
5 June	1110	9.57	0.86	0.041	0.045	0.0051

Γ_1 is the potential temperature gradient in the air below 100 m,

Γ_2 is the gradient of the potential temperature above this height.

The results from the simulations are shown in Fig. 4; for comparison the measured values of the IBL is also shown. Good agreement between the results of the simulations and measurements is seen for all three days.

Acknowledgements

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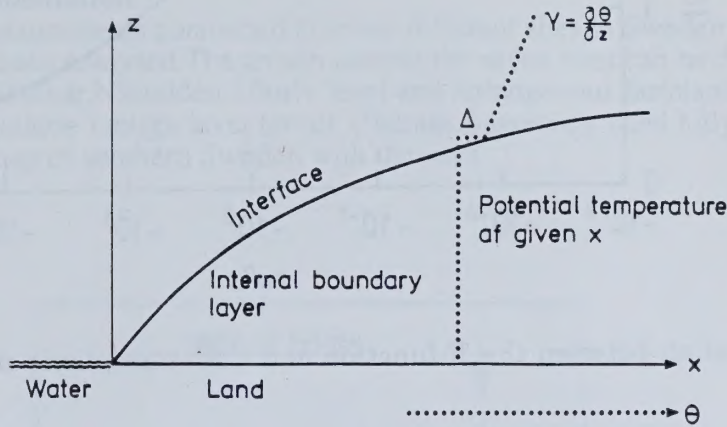


Fig. 1 Schematic illustration of the physical system that forms the basis of the model.

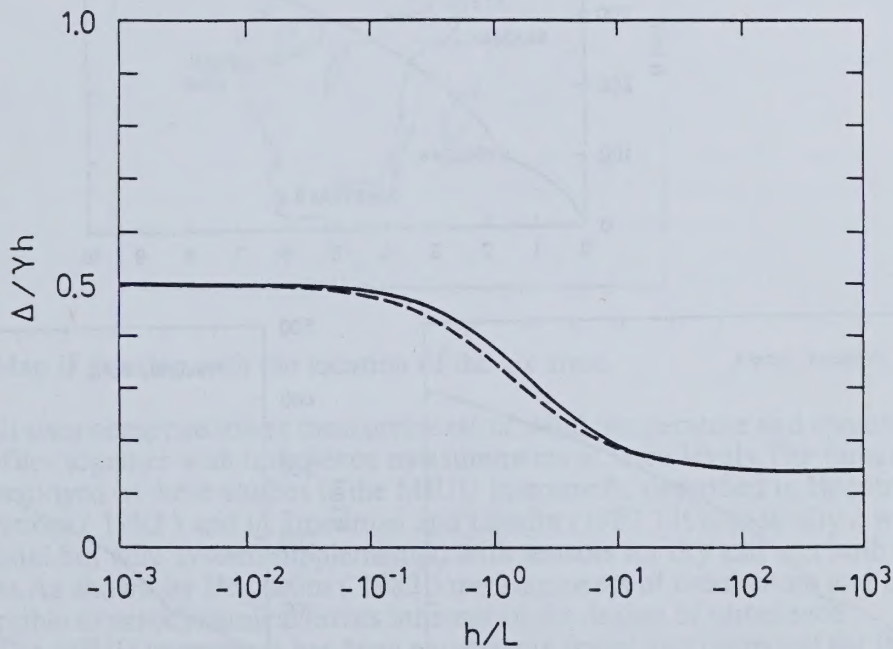


Fig. 2 Dimensionless temperature difference across the inversion, $\Delta/\gamma h$, as function of atmospheric stability h/L . The full line shows the analytical solution; the dashed line its approximation, Eq.(11).

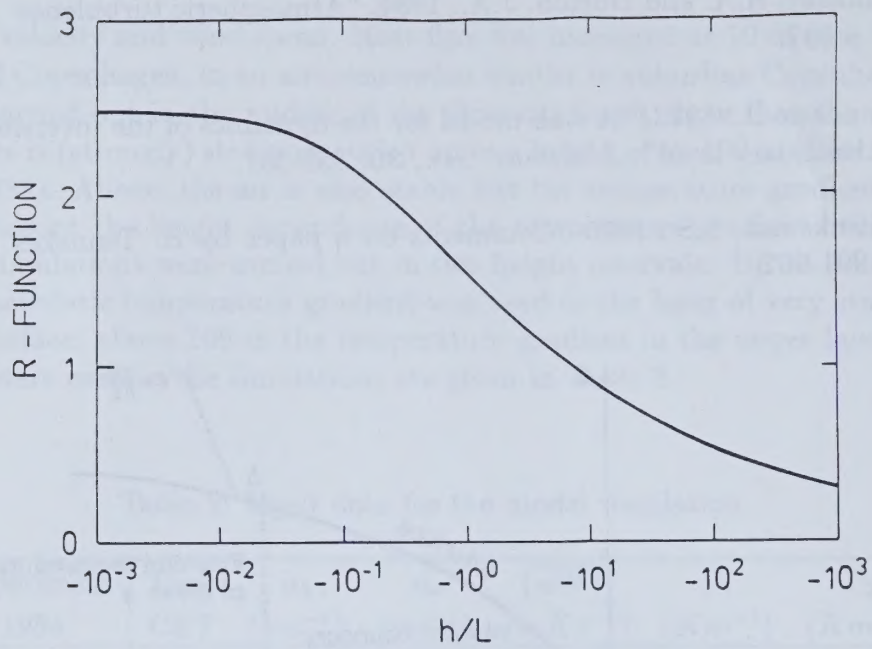


Fig. 3 Relation between the R -function and the atmospheric stability parameter h/L .

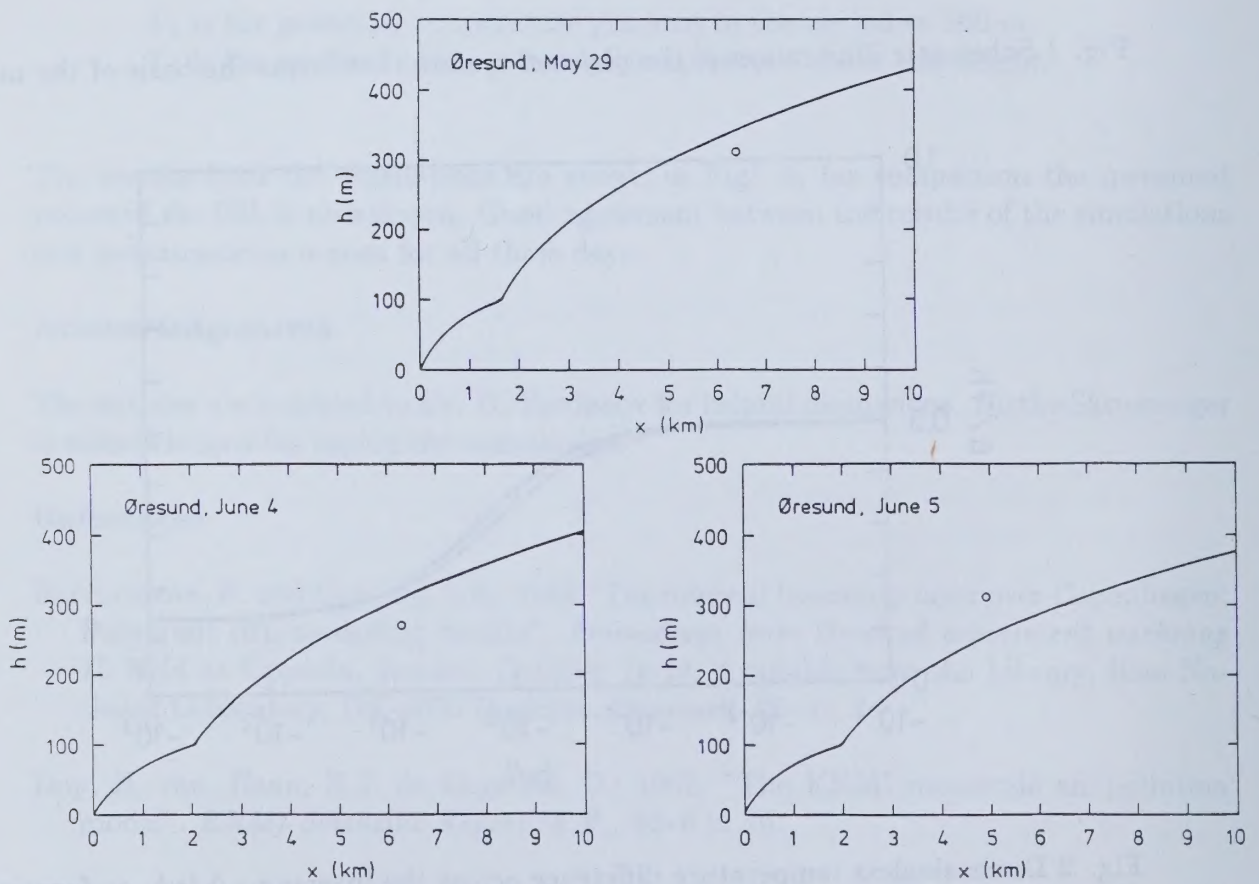


Fig. 4 Simulations of the growth of the internal boundary layer. Also shown are the measured values.

Observations of turbulence structure in stable boundary layers over different kinds of terrain - including the Öresund area.

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1. Sites and instrumentation.

Boundary layer measurements conducted at seven different sites in Sweden during stable stratification have been analysed. The terrain around the seven sites can be divided into four groups, open sea (Nässkärs, Näsudden), fairly level and homogenous farmland (Marsta, Lövsta, Näsudden), rough level terrain (Jädraås, Barsebäck) and hilly farmland (Vänernborg). Figure 1 shows a map of southern Sweden with the sites

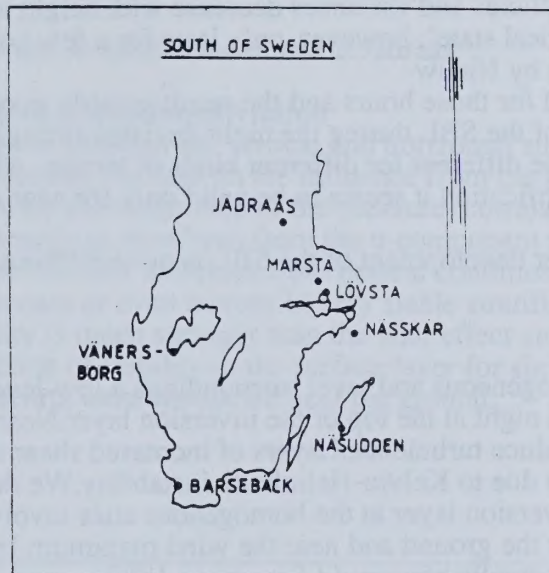


Figure 1. Map of Sweden with the location of the six sites.

The data at all sites comprise tower measurements of wind, temperature and sometimes humidity profiles together with turbulence measurements at three levels. The turbulence instrument employed in these studies is the MIUU instrument, described in Högström et al. (1980), Högström (1982) and in Smedman and Lundin (1987). It is basically a wind vane based three axial hot wire system supplemented with sensors for dry and wet bulb measurements. As shown by Högström (1982) measurements of momentum and heat flux are highly susceptible to aerodynamical errors inherent in the design of turbulence instruments. The MIUU instrument has been extensively tested and corrected for this type of error.

The heights of the different towers and the levels of measurements are given in Table 1. Doppler acoustic sounders together with a tethered balloon sounding system and radiosonde ascents gave information of the wind and temperature profiles up to 300-500 m.

Table 1.Measuring heights at the sites.

Sites	Js	Ma	La	Nr	Nm	Vg	Bk
Heights of tower	50	30	24	31	145	24	100
Levels of prof. mea.	10	6	6	4	6	5	6
Heights of turb.	29,38 47	2,6 30	2,6 14	8,31	10,77 135	2,15 24	10,40 95

2.The development of the stable boundary layer.

The development of the stable boundary layer during the night is similar at all sites. Shortly after sunset a rapid transition to a stable layer occurs and the height of the turbulent layer is about the same as the height of the inversion layer. During the evening the height of the inversion layer increases whereas the height of the turbulent stable boundary layer (SBL) decreases. In the SBL the fluxes and variances decrease with height and become negligible at the top of the SBL. This 'ideal state', however, only lasts for a few hours. The local scaling similarity theory proposed by Nieuwstadt (1984) can be tested for those hours and the result is fairly good.

The further development of the SBL during the night deviates strongly from the 'ideal state' described above and will be different for different kinds of terrain. Although the local scaling theory assumes z-less stratification it seems to be valid only for near neutral to moderately stable situations.

During the night the further development of the SBL is quite different at the sites depending on topography.

2.1 Homogeneous sites.

At the four sites with homogeneous and level surroundings a low-level jet will very frequently form during the night at the top of the inversion layer. Near this wind maximum the increased shear tends to induce turbulence. Layers of increased shear can also produce organized waves, probably due to Kelvin-Helmholtz instability. We thus conclude that the vertical structure of the inversion layer at the homogenous sites involve two layers of significant turbulence; near the ground and near the wind maximum. In one case separate layers of turbulence were actually observed (Smedman, 1988).

2.2 Sites with rough terrain.

At the sites situated in more complex terrain the development is quite different. No low-level jets are observed at the top of the inversion layer, except in cases with very low wind speed. The turbulence seems to be confined to the adjacent surface and there are no indications of wave activity in the low frequency range of the energy and cross spectra.

3.Turbulence structure in the SBL.

In this investigation one aim has been to look for turbulence parameters valid in the SBL at all sites regardless of terrain.

3.1 The form of the spectral curve.

At all sites the form of the curves of normalized spectra and cospectra show the usual behaviour found over smooth terrain. Figure 2 shows normalized w-spectra at Jädraås (the forest) The solid curve is the expression proposed by Kaimal (1973).

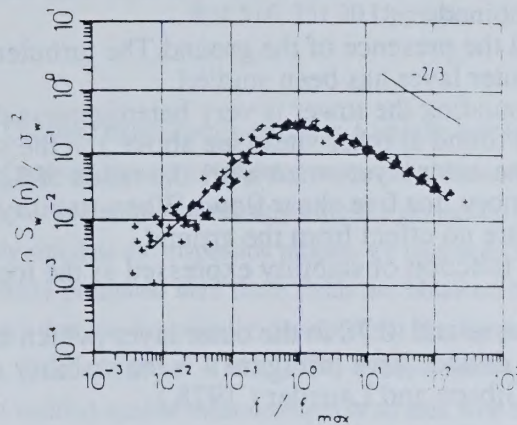


Figure 2. Normalized w-spectra measured at Jädraås.

3.2 Vertical to horizontal standard deviation

The variation of the ratio between the vertical and horizontal standard deviation in stable air will depend on stratification but also on the influence from the ground. The presence of a wall in the flow will decrease the magnitude of the pressure-correlation term in the turbulence energy budget. Less energy is transferred from the u-component to the w-component. Thus in the surface layer when stability increases from neutral condition the wall effect decreases and σ_w/σ_u may remain constant or even increase. In very stable stratification, however, the dependence on stability is much stronger than the wall effect and σ_w/σ_u will decrease. Figure 3 shows σ_w/σ_u as a function of stability in the surface layer for six of the sites. At the forest site Jädraås no measurements were conducted near the ground.

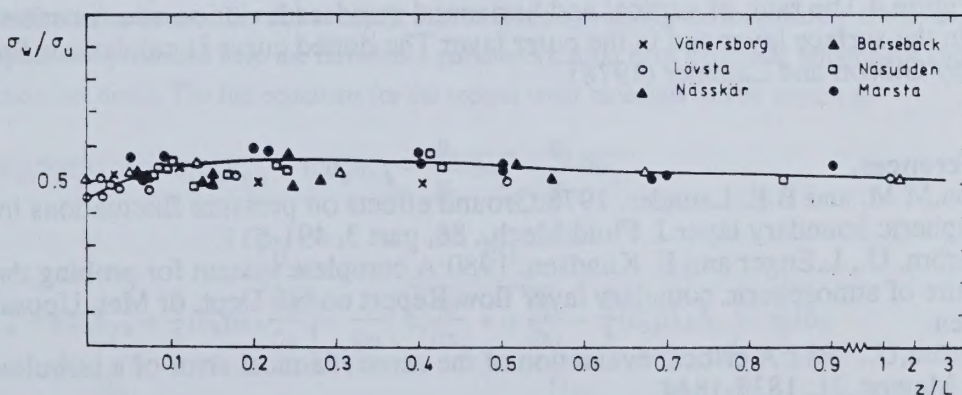


Figure 3. Vertical to horizontal standard deviation as a function of stability at six sites.

3.3 Vertical to horizontal heat flux.

The above discussion can also be applied to the ratio of the vertical and horizontal heatflux. Again stratification and the wall effect will compensate each other over a large frequency range. When stability exceeds z/L 1 the stratification effect dominates over the wall effect. The site (Vänersborg) with slightly sloping terrain deviates in this case from the six other sites with level terrain.

4. The Barsebäck site.

Measurements taken at the Barsebäck site in the Öresund area has been looked into in more detail concerning the combined effect of stratification and the presence of the ground. The turbulence structure both in the surface layer and in the outer layer has been studied.

Although the terrain surrounding the tower is very heterogenous the turbulence parameters studied, confirm the result found at other sites (see above) in the surface layer.

For neutral condition in the outer layer when $z/h \gg 1$ the ratios of σ_w/σ_u and $w\theta/u\theta$ approach the values found in the laboratory for free shear flows. When stability increases the different ratios decrease and there are no effect from the ground.

Figure 4 shows σ_w/σ_u as a function of stability expressed as the local Monin-Obukhov length (L_z). At neutral condition

$\sigma_w/\sigma_u = 0.45$ in the surface layer and 0.70 in the outer layer, which are the values found in wind tunnel measurements. The dotted curve in Figure 4 is the stability variation for free shear flow calculated by Gibson and Launder (1978)

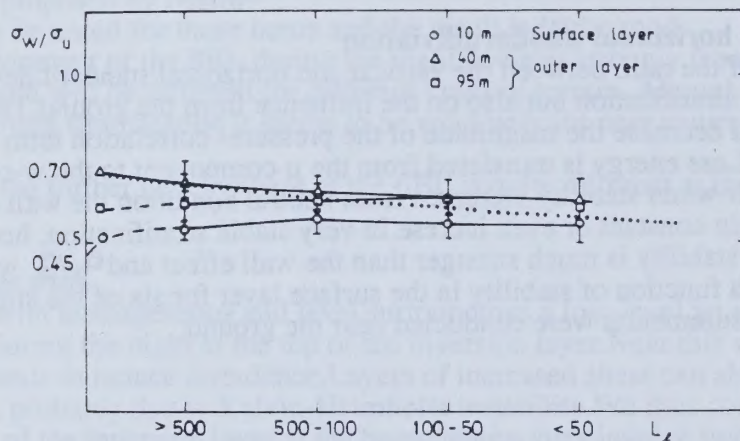


Figure 4. The ratio of vertical and horizontal standard deviation as a function of stability both in the surface layer and in the outer layer. The dotted curve is calculations for free shear flow by Gibson and Launder (1978).

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SIMULATION OF THE ÖRESUND-DATABANK TRACER EXPERIMENTS

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1. Introduction.

The tracer data from the Öresund experiments provides a unique test bench for dispersion models aimed at simulation of dispersion of passive tracers in inhomogeneous conditions. Since the area is characterized by large contrasts in surface conditions we can expect strongly differing dispersion conditions along the paths of the tracer plumes. In order to properly simulate the dispersion process we thus need a detailed description of the turbulence fields in the area. In the study presented here these fields are obtained from a three-dimensional version of a second-order closure model that has been under development during the last six years at the Department of Meteorology in Uppsala, Sweden (Enger, 1983, 1984, 1986; Grandin, 1984; Tjernström, 1987, 1988). The model has previously been verified against measurements in an area with noticeable small scale topography, and has been found to give good results (Tjernström 1987; Enger 1988). For the simulations to be presented an improved closure for the pressure redistribution terms has been included, following Gibson and Launder (1978), hereafter referred to as G&L.

A short description of the basic model assumptions will be presented together with a discussion of their implications, followed by a comparison of observed and calculated turbulence fields.

Based on the turbulence parameterization in the meso- γ -scale model a simplified second-order closure dispersion model is formulated and some preliminary calculations of tracer plumes described.

2. Meso- γ -scale model formulation.

The model consists of the Boussinesq approximated forms of the equations for mean quantities and second order moments. Pressure is separated into a synoptic part, assumed to be in geostrophic balance, and a mesoscale perturbation. The mesoscale pressure perturbation is assumed to be zero on model top and diagnosed by integrating the hydrostatic assumption from the model top downwards. Vertical velocity has been diagnosed from the anelastic form of the equation of continuity. The model version used in the present simulations does not include moisture or radiation. The diurnal cycle is prescribed by varying the temperature at the lower boundary. For the application presented here the turbulence parameterization is of particular importance and will be dealt with in some more detail. The full equations for the second order moments can be written as

$$\begin{aligned} \frac{\partial \overline{u_i u_j}}{\partial t} + \overline{u_k (u_i u_j)_{;k}} = & - \overline{u_i u_k u_{j;k}} - \overline{u_j u_k u_{i;k}} - \frac{g_i}{\theta} \overline{\theta u_j} - \frac{g_j}{\theta} \overline{\theta u_i} \\ & - \left(\overline{u_j u_k u_{i;k}} + \overline{u_i u_k u_{j;k}} + \frac{2}{3} (\overline{u_k p})_{;k} \frac{\delta_{ij}}{\rho_0} \right) - \frac{1}{\rho_0} \left(\overline{u_j \frac{\partial p}{\partial x_i}} + \overline{u_i \frac{\partial p}{\partial x_j}} - \frac{2}{3} (\overline{u_k p})_{;k} \delta_{ij} \right) - \frac{2}{3} \epsilon \delta_{ij} \end{aligned} \quad (1)$$

$$\begin{aligned} \frac{\partial \overline{\theta u_i}}{\partial t} + \overline{u_j (\theta u_i)_{;j}} = & - \overline{u_i u_j \frac{\partial \theta}{\partial x_j}} - \overline{\theta u_j u_{i;j}} - \frac{g_i}{\theta} \overline{\theta^2} - (\overline{u_i u_j \theta})_{;j} - \frac{1}{\rho_0} \overline{\theta \frac{\partial p}{\partial x_i}} \end{aligned} \quad (2)$$

$$\frac{\partial \bar{\theta}^2}{\partial t} + \bar{u}_j \frac{\partial \bar{\theta}^2}{\partial x_j} = -2\bar{\theta} \bar{u}_j \frac{\partial \bar{\theta}}{\partial x_j} - (\bar{u}_j \bar{\theta}^2)_{,j} - 2\varepsilon_\theta \quad (3)$$

D_j

For the simulations to be presented a level 2.5 version has been used. Thus for second order moments turbulent transport and advection are kept only in the equation for turbulent energy, i.e. the trace of (1). A standard gradient diffusion parameterization is used for D_{ij} . Dissipation is taken as the quotient between the dissipated quantity and a turbulent timescale. The redistribution terms, Π_{ij} and Π_i , are of fundamental importance in connection with dispersion applications since they determine the partition of turbulent kinetic energy among the velocity variances. Denoting nonlinear turbulence interactions by Π_{ij}^{TT} , mean shear - turbulence interactions by Π_{ij}^S and buoyancy - turbulence interactions by Π_{ij}^B , an often used closure for these terms, based on work by e.g. Launder, Reese and Rodi (1975) and Lumley (1979), is

$$\Pi_{ij}^{TT} = C_1 \varepsilon b_{ij} \quad (4)$$

$$\Pi_{ij}^S = -2q^2 \left\{ \frac{1}{5} S_{ij} + \alpha_1 \left(S_{ik} b_{kj} + S_{jk} b_{ki} - \frac{2}{3} S_{kl} b_{lk} \delta_{ij} \right) + \alpha_2 \left(R_{ik} b_{kj} + R_{jk} b_{ki} \right) \right\} \quad (5)$$

$$\Pi_{ij}^B = -\frac{3}{10} \frac{g_k}{\theta_0} \left(\bar{\theta} u_k \delta_{kj} + \bar{\theta} u_j \delta_{ki} - \frac{2}{3} \bar{\theta} u_k \delta_{ij} \right) \quad (6)$$

$$\Pi_i^{TT} = C_1 \theta \frac{\varepsilon}{q^2} \bar{\theta} u_i \quad (7)$$

$$\Pi_i^S = -\frac{4}{5} \bar{\theta} u_j \bar{u}_{i,j} + \frac{1}{5} \bar{\theta} u_j \bar{u}_{j,i} \quad (8)$$

$$\Pi_i^B = -\frac{1}{3} \frac{g_i}{\theta_0} \bar{\theta}^2 \quad (9)$$

Where the b_{ij} , S_{ij} and R_{ij} denotes

$$b_{ij} = \frac{\bar{u}_i \bar{u}_j}{q^2} - \frac{1}{3} \delta_{ij}, \quad S_{ij} = \frac{1}{2} \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} \right), \quad R_{ij} = \frac{1}{2} \left(\frac{\partial \bar{u}_i}{\partial x_j} - \frac{\partial \bar{u}_j}{\partial x_i} \right) \quad (10)$$

In earlier versions of the model a less complicated closure for the rapid terms, i.e. shear and buoyancy terms, was used in which only the first term on the right hand side of Π_{ij}^S was kept, with the invariant constant replaced by a tunable constant. This closure was based on works by Mellor (1973) and Mellor and Yamada (1974). In this paper the closure of the pressure redistribution terms is changed so as to include wall effects. Several authors have reported problems with realizability using the above formulation for the rapid terms in the heat-flux equations. For example Wyngaard (1975) allowed the constant in the Π_i^B term to be stability dependent. These problems are discussed in some detail by Zeman and Lumley (1979). Some of these problems are probably due to the neglect of wall effects.

In the closure used by Mellor (1973) and Mellor and Yamada (1974) the closure constants were matched against wind tunnel data with quite high $\bar{u}w/q^2$ - values, about -0.16 in their papers. In the atmospheric surface layer this ratio is much smaller. Enger (1986) used -0.11, based on the Kansas measurements. Panofsky and Dutton (1984) presents a collection of data giving a value of -0.09. Forcing the closure proposed by Mellor and Yamada (1974) to this stress level results in an unrealistic behaviour of the equation system; the well defined critical Richardson number is lost and the equations can support a non-zero temperature variance although the turbulent kinetic energy is zero, further the ratio between turbulent exchange coefficients for heat and momentum becomes less than 1 for unstable stratification and greater than 1 for stable stratification, in contradiction with measurements,

e.g. Högström (1988). Inclusion of the full closure (4) - (10), using $\alpha_1=\alpha_2=0.3$ for realizability (Zeman, 1981), does not remedy this problem. The resultant exchange coefficients becomes rather close to each other but are still erroneous, as well as the behaviour at strong stability. In addition the lateral and vertical velocity variance components are predicted as equal, an unwanted effect if the simulated fields are to be used for dispersion calculations. We can thus conclude that for a successful description of atmospheric boundary layer turbulence wall effects on the pressure redistribution terms must be considered. For the simulations to be presented the parameterization of wall effects suggested by G&L has therefore been included. Their suggestion for wall effects on the redistribution terms in the Reynolds tensor is

$$\Pi_{ij}^{TTW} = -D_1 \frac{2\varepsilon}{q^2} \left(\overline{u_k u_m} n_k n_m \delta_{ij} - \frac{3}{2} \overline{u_k u_i} n_k n_j - \frac{3}{2} \overline{u_k u_j} n_k n_i \right) f \left(\frac{1}{n_i r_i} \right) \quad (11)$$

$$\Pi_{ij}^{SW} = -D_2 \left(W_{km}^S n_k n_m \delta_{ij} - \frac{3}{2} W_{ik}^S n_k n_j - \frac{3}{2} W_{jk}^S n_k n_i \right) f \left(\frac{1}{n_i r_i} \right) \quad (12)$$

$$\Pi_{ij}^{BW} = -D_3 \left(W_{km}^B n_k n_m \delta_{ij} - \frac{3}{2} W_{ik}^B n_k n_j - \frac{3}{2} W_{jk}^B n_k n_i \right) f \left(\frac{1}{n_i r_i} \right) \quad (13)$$

Where W_{km}^S and W_{km}^B denotes

$$W_{km}^S = -C_2 \left(P_{km} - \frac{2}{3} P \delta_{km} \right), \quad W_{km}^B = -C_3 \left(B_{km} - \frac{2}{3} B \delta_{km} \right) \quad (14)$$

For the redistribution terms in the heat flux equations the proposed form is

$$\Pi_i^W = D_{1\theta} \frac{2\varepsilon}{q^2} \overline{u_k \theta} n_k f \left(\frac{1}{n_i r_i} \right) + D_{2\theta} C_{2\theta} \overline{u_i \theta} \frac{\partial \overline{u_k}}{\partial x_k} n_k f \left(\frac{1}{n_i r_i} \right) + D_{3\theta} C_{3\theta} \frac{g_k}{\theta_0} \overline{\theta^2} n_k f \left(\frac{1}{n_i r_i} \right) \quad (15)$$

Formulated in this manner the wall terms counteracts the 'ordinary' redistribution terms and acts so as to produce larger anisotropy. In the paper by G&L the wall terms are weighted by the function f , taken as $f = 1 / \kappa z$, where l is a length scale characterizing the energy containing eddies. In this paper we take l equal to the master length scale. It is however found, that even with the inclusion of wall effects in this manner, it is not possible to describe the low stress levels given by e.g. Panofsky and Dutton (1984), and still keep the quotient between exchange coefficients and the Richardson number well behaved. G&L quotes a value of $\overline{uw} / q^2$ equal to -0.13 for near wall turbulence. For the simulations to be presented we have adopted this value, which is only 15% higher than what has been used with the model before, (Enger, 1986). Away from the surface we follow the $\overline{u_i u_j} / q^2$ - values used by G&L. Using $\alpha_1=\alpha_2=0.3$ and dropping the second term in (15) makes the closure for the redistribution terms identical to the one used by G&L. With wall effects included in this manner a reasonable behaviour of the exchange coefficients is obtained, as well as a well defined critical Richardson number for the set of equations. The turbulence time scales appearing in the above parameterizations have been taken as proportional to the ratio between a master length scale and the square root of twice the turbulent kinetic energy. Thus all turbulence time scales have the same vertical structure in the model. Further details on the closure scheme, closure constants and length scales can be found in Andréén (1988).

3. Evaluation of turbulence fields.

The data bank provides turbulence measurements mainly at the Barsebäck tower, but an estimate of turbulent kinetic energy can also be obtained at 115 m height at the Gladsaxe tower. In addition the airborne measurements provide a relative measure of turbulence through the given R-values obtained during flight traverses in the area. According to Sievertsen (1986) the R-values are related to the dissipation of turbulent kinetic energy as

$$R \propto \varepsilon^{1/3} \rightarrow R \propto \frac{q}{(C_\varepsilon l)^{1/3}} \quad (16)$$

using the model parameterisation of dissipation.

The R-values thus provides a way of judging relative levels of dissipation of turbulent kinetic energy in the area, by scaling 'modelled' R-values to the measured value at a point on a flight traverse.

A key parameter in this type of model is the turbulent length scale, determining through dissipation the level of turbulent kinetic energy in the simulations. Profiles of measured and calculated turbulent kinetic energy at the Barsebäck tower for 4, 5 (offshore flow) and 12 of June (onshore flow), displayed in fig. 1a-c, shows that the model gives a satisfying result. The slightly low values at 11.00 and onwards on 4 June is the result of an increase in synoptic forcing, not covered by the present model formulation. A critical test of the length scale scheme can be obtained through the above mentioned R-values. The results for westerly winds shown in fig.2 and the results for easterly winds in fig.3 gives an encouraging picture. The model seems to give the proper decay of turbulence as the flow passes from land over to the smoother and colder sea surface. The rapid growth of the R-values on the western side of the strait, shown in fig. 2 also indicates that the growth of the internal boundary layer over the downstream land area is correctly predicted. The tower at Barsebäck is situated on a small peninsula some 200m from the shore and about 500m from the nuclear power station buildings. Although we can expect some very local influences on the measurements, not covered by the gridresolution of 3.5 km, an attempt has been made to compare measured and calculated quotients between second order moments. Fig. 4a shows quotients between longitudinal and vertical velocity variances and between longitudinal and vertical turbulent heatfluxes for 4 June, and fig. 4b the same for 12 June. Except for the 10m level there is a good agreement for the heatflux quotients. The profile shapes for the velocity variance quotients looks reasonable, but the model seems to underestimate the vertical variance. This would indicate that the wall effect is too heavily weighted. Further evaluation at a more ideal site should be made before any definite conclusions are drawn.

4. Dispersion model formulation.

Since the fields of turbulence and mean quantities are strongly inhomogeneous we cannot expect the tracer plumes to be well approximated by Gaussian distributions. In order to investigate the detailed downstream development of the plumes a simplified second order closure model has been formulated. The model is based on the same closure assumptions as the turbulence model. By assuming that advection and turbulent transport terms only has a limited effect on the second order moments we can extract diagnostic expressions for these moments. In the terminology of Mellor and Yamada this would then be a level one second order closure dispersion model. With this approach the conservation equation for mean concentration is obtained as

$$\begin{aligned} \frac{\partial \bar{\chi}}{\partial t} + U^r \frac{\partial \bar{\chi}}{\partial r} + U^\theta \frac{\partial \bar{\chi}}{\partial \theta} + U^\sigma \frac{\partial \bar{\chi}}{\partial \sigma} = \frac{1}{\sqrt{G}} \frac{\partial}{\partial r} \left(\sqrt{G} \left(D^{11} \frac{\partial \bar{\chi}}{\partial r} + D^{12} \frac{\partial \bar{\chi}}{\partial \theta} + D^{13} \frac{\partial \bar{\chi}}{\partial \sigma} \right) \right) + \\ \frac{1}{\sqrt{G}} \frac{\partial}{\partial \theta} \left(\sqrt{G} \left(D^{21} \frac{\partial \bar{\chi}}{\partial r} + D^{22} \frac{\partial \bar{\chi}}{\partial \theta} + D^{23} \frac{\partial \bar{\chi}}{\partial \sigma} \right) \right) + \frac{1}{\sqrt{G}} \frac{\partial}{\partial \sigma} \left(\sqrt{G} \left(D^{31} \frac{\partial \bar{\chi}}{\partial r} + D^{32} \frac{\partial \bar{\chi}}{\partial \theta} + D^{33} \frac{\partial \bar{\chi}}{\partial \sigma} \right) \right) \end{aligned} \quad (17)$$

where a cylindrical polar terrain following system have been used in order to resolve the plume close to the source. G denotes the square root of the determinant of the Jacobian of the transformation. $D^{11} - D^{33}$ depends on the closure assumptions and length and velocity scales of the plume, and thus are dynamically coupled to the plume growth. The equation is solved by splitting, using finite element solutions with linear basis functions for the advection terms, and a Crank-Nicholson scheme for the diffusion steps.

5 Preliminary plume calculations.

For the preliminary simulations presented here the plume scales are taken as equal to their analogous turbulence scales. Fig. 5 displays a sequence of downstream cross-sections of a plume on 4 June. The plume is strongly distorted by the winddirection shear in the lower layers over the strait. The decay of turbulence over the strait results in a very small vertical spreading of the plume, with the result that the elevated maximum is still present

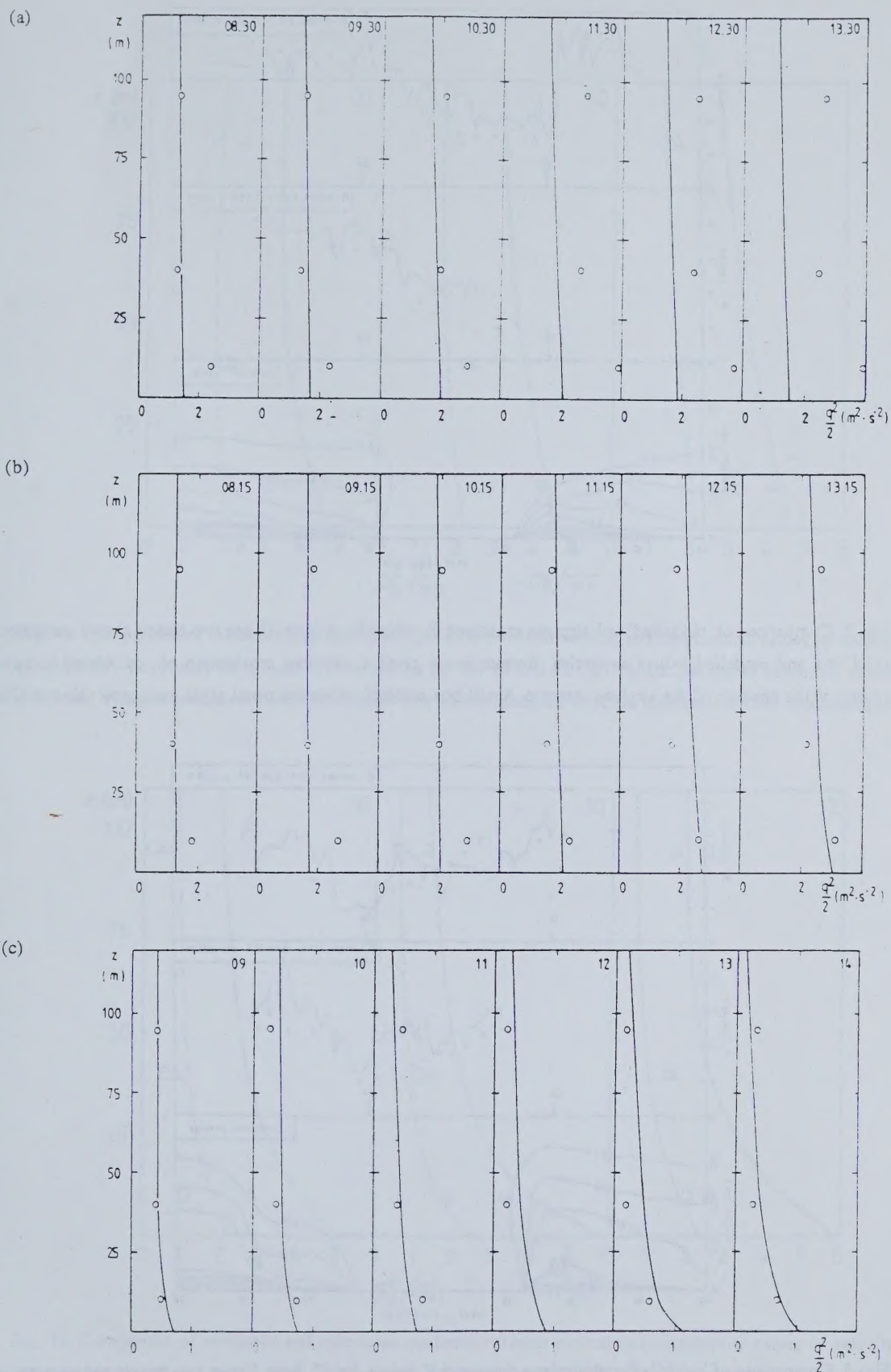
as the plume reaches the shoreline 20 km downstream of the source. As the plume encounters the internal boundary layer on the Danish side of the strait it is rapidly mixed in the vertical, and an increase in concentration at ground level is obtained, Fig. 6.

Judging from preliminary runs, the dispersion model seems to be a useful tool for the study of plumes spreading in inhomogeneous turbulence fields.

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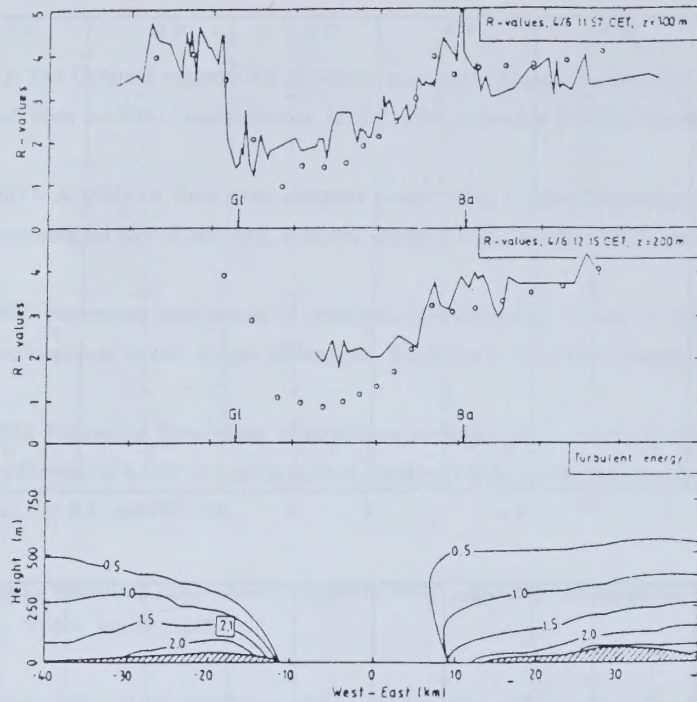


Fig. 2. Comparison of 'modelled' and airplane measured R-values for 4 June. Upper two panels shows measured values as solid line and modelled values as circles. Bottom panel gives a west-east cross-section of calculated turbulent kinetic energy at the position of the airplane traverse. Small box inserted in bottom panel gives measured value at Gladsaxe.

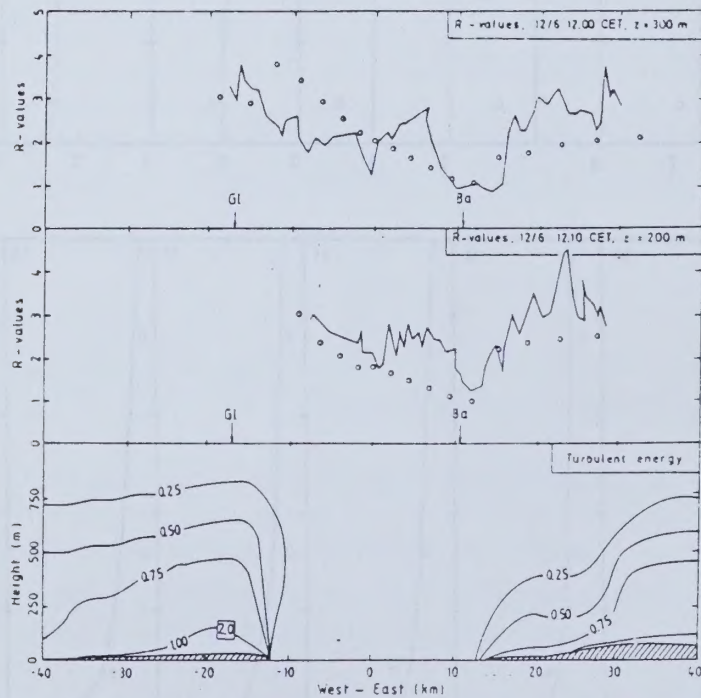


Fig. 3. Comparison of 'modelled' and airplane measured R-values for 12 June. Upper two panels shows measured values as solid line and modelled values as circles. Bottom panel gives a west-east cross-section of calculated turbulent kinetic energy at the position of the airplane traverse. Small box inserted in bottom panel gives measured value at Gladsaxe.

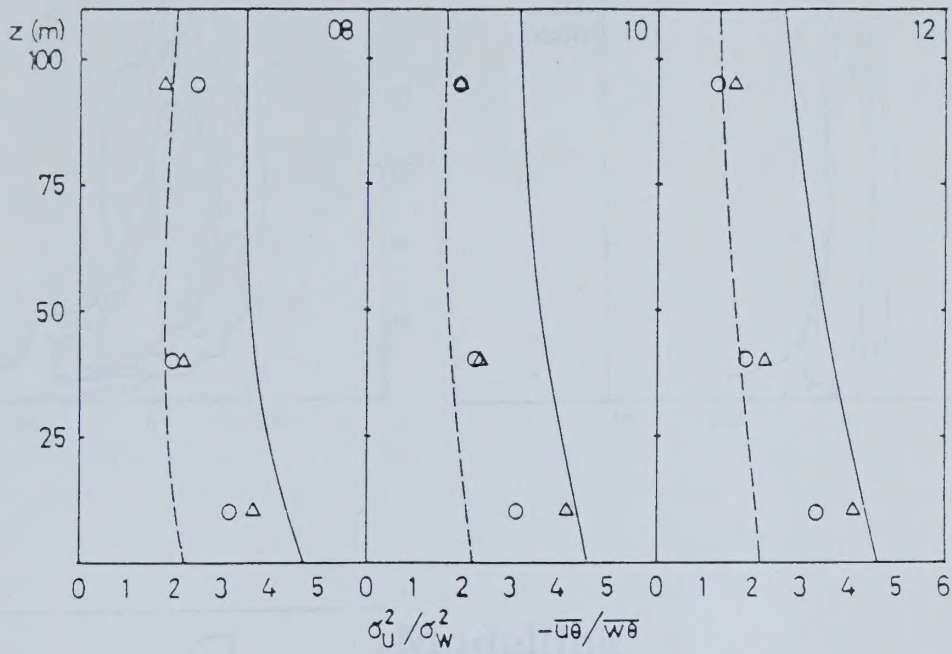


Fig. 4a. Comparison of measured and calculated quotients between vertical and longitudinal values of velocity variances and turbulent heatfluxes for 4 June. Values shown for 08, 10 and 12 CET. Solid line shows modelled velocity variance quotients, triangles measured quotients. Dashed line shows modelled heatflux quotients, circles measured quotients.

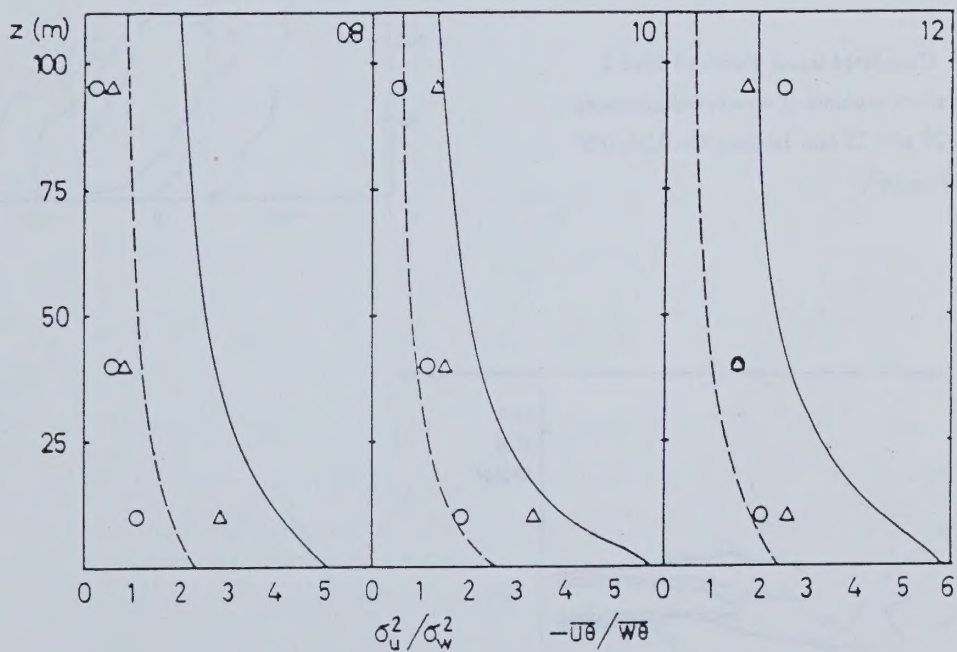


Fig. 4b. Comparison of measured and calculated quotients between vertical and longitudinal values of velocity variances and turbulent heatfluxes for 12 June. Values shown for 08, 10 and 12 CET. Solid line shows modelled velocity variance quotients, triangles measured quotients. Dashed line shows modelled heatflux quotients, circles measured quotients.

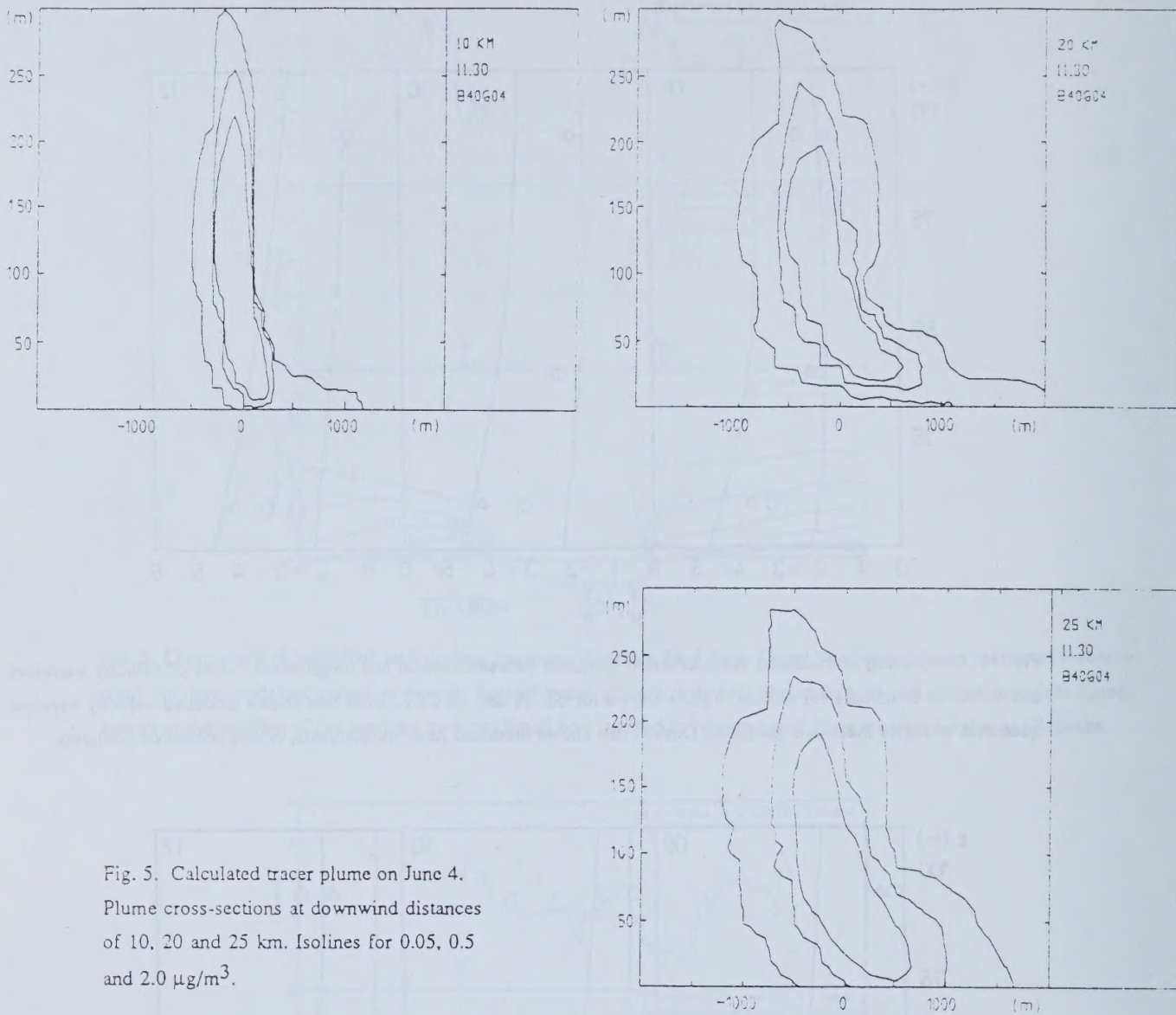


Fig. 5. Calculated tracer plume on June 4.
Plume cross-sections at downwind distances
of 10, 20 and 25 km. Isolines for 0.05, 0.5
and $2.0 \mu\text{g}/\text{m}^3$.

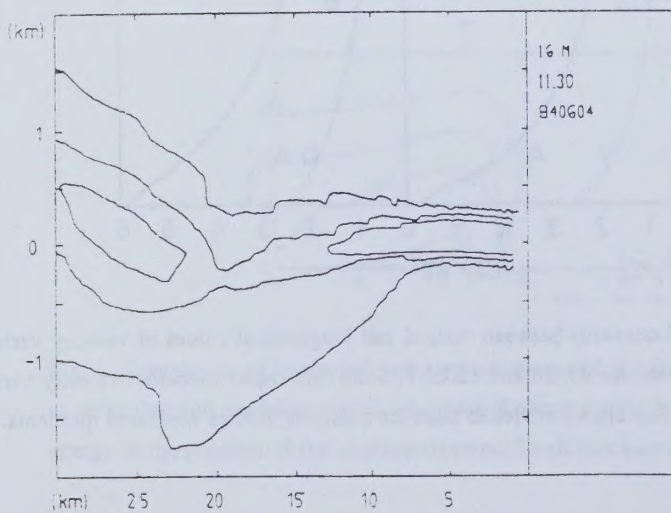


Fig. 6. Calculated tracer plume on June 4.
Horizontal cross-section at a height of 15m.
Isolines for 0.05, 0.5 and $2.0 \mu\text{g}/\text{m}^3$.

Modelling

A THREE-DIMENSIONAL NUMERICAL STUDY OF POLLUTANT DISPERSION
IN A COASTAL VALLEY

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INTRODUCTION

The Latrobe Valley, about 120 km southeast of Melbourne, lies approximately parallel to the coastline of southeast Australia. It is typically 15 km wide and oriented east-northeast/west-southwest from Yallourn to Rosedale before widening to its mouth near the east coast. The western section between Warragul and Yallourn lies east-west and is narrower (above 8 km). The Great Dividing Range north of the Valley rises to 2000 m above sea level, while the height of the Strzelecki Range to the south averages about 500 m.

Large brown coal deposits in the area are used by the State Electricity Commission of Victoria to fuel 5 power stations in the Valley, while industries such as Australian Paper Mills are also sited in the region. For air quality purposes, the application of Gaussian plume models, in which uniform winds in space and time are assumed, is severely restricted by the existence of mesoscale circulations generated by the complex topography.

In this paper, an air quality system is described in which a three-dimensional numerical model, capable of simulating mesoscale effects in complex terrain, is used to predict wind fields over the period of interest. Wind and turbulence fields from the model are used to transport and diffuse pollutants from any number of sources.

OBSERVATIONS

On 8 March 1985, an anticyclone centred off the east Australian coast produced light northeasterly flow over the Latrobe Valley. This flow was evident on the MSL, 850 hPa and 700 hPa charts. This synoptic regime influenced the Latrobe Valley region until 12 March 1985.

During this period, early morning wind soundings in the Valley showed weak low-level flow with an easterly component. Above 400 to 600 m, and to a height of at least 2500 m, stronger flow with a westerly component was observed, in direct contrast to the synoptic pressure gradient at these levels. As the morning progressed, heating of the land resulted in the convective mixing of the westerlies to the surface.

The diurnal heating also produced sea breezes and upslope flows. In the morning, sea breeze circulations were set up along the eastern and southern coastlines. During the day, these two sea breezes propagated inland and intersected in late afternoon near Morwell, 80 km up the Valley from the east coast. The south coast sea breeze came over a saddle in the Strzelecki Range. By early evening, easterly flow dominated the region with strong terrain effects (channelling) evident where the Valley begins to narrow. This low-level flow remained strong throughout the night, in contrast to the almost calm conditions which prevailed near the coastline. Next morning the westerly winds progressively replaced the easterlies from above and the cycle was repeated.

WIND MODELLING

The Colorado State University Mesoscale Model (Mahrer and Pielke 1977, McNider and Pielke 1981), modified to include radiative lateral boundary conditions and an alternate PBL height scheme, has been used to model the wind regime in the Latrobe Valley region. A domain size of 50 x 50 x 22 gridpoints with a grid spacing of 5 km was used. Other parameter values used to initialize the model are based on conditions representative of 8 March 1985. In these simulations the geostrophic wind was considered to be negligible. It has been found that in order to simulate this day successfully, the model must be run for two days. On the first day the diurnal processes discussed earlier produce sea breezes and upslope flows. During the following night, it appears that an inertial oscillation of the return flows results in the production of westerlies at upper levels. On the second day these westerlies are convectively mixed to the surface in the western section of the valley, prior to the onset of the sea breeze. The westerly winds also retard the onset of the east coast sea breeze on the second day.

Comparison of model results with observations on 8 March 1985 shows quite good agreement. During the afternoon, the model accurately predicts the inland extent of both sea breezes and the strength of the westerlies ahead of the sea breeze. At night, the much stronger winds observed in the western section of the Valley (than in the eastern) are also well simulated.

DISPERSION MODELLING

Pollutant dispersion is investigated using a Lagrangian scheme in which individual particles are transported and diffused according to wind and turbulence fields provided by the mesoscale model.

The relevant fields are supplied on a three-dimensional grid every 10 minutes, although this time interval is variable. Particles, which are assumed to be inert and neutrally buoyant, can be released at any rate from any number of locations in the horizontal and vertical and are moved with a 20 second timestep. The basis of the diffusion component is the "conditioned particle" scheme in which a turbulent velocity fluctuation is linearly related to the fluctuation at an earlier time through an autocorrelation coefficient. Such fluctuations also include a random component which is taken to be Gaussian. A review of this approach to turbulent dispersion modelling can be found in Sawford (1985).

It follows that the strongly-sheared and non-steady wind fields observed, and modelled, on 8 March should produce non-Gaussian plume behaviour. As an illustration, numerical experiments were carried out in which particles were released from 0600 local time on the second day at a rate of one per minute from sources located at a) Morwell Power Station and b) the Australian Paper Mills (APM) site 10 km northeast of the Morwell source. The plume rise height was specified as 550 m AGL for Morwell and 100 m for APM. The low-level shear discussed earlier results in early morning emissions from the two sources moving in significantly different directions from each other, although by noon all emissions proceeded towards the northeast. By midday, early emissions from the Morwell site have been taken southward by upslope winds on the Strzelecki Ranges whereas APM emissions were moving northward in an upslope wind system on the opposite side of the Valley. This example illustrates both the necessity of allowing for vertical and horizontal wind shear and also the significant influence of effective stack height on plume trajectory under such conditions.

By 1800 local time, emissions from both sources have been advected back up the Valley by the east coast sea breeze with particles tending to gather at its leading edge and at the boundary between the two sea breezes. Clean air is evident at low levels further behind the leading edge, as convective mixing has ceased by this time and the plume remains aloft. A rise in surface concentration levels on sea breeze passage followed by cleaner air, as predicted, is often observed at stations west of the Morwell industrial region. To the east towards the coastline, the sea breeze is observed and predicted to transport clean maritime air up the Valley.

For a fictitious source located near the coastline, the sea breeze arrives before noon and particles are convectively mixed throughout the boundary layer and advected inland during the day. In this situation, the sea breeze does not sweep out the Valley with fresh maritime air, but actually causes pollutant levels to increase.

CONCLUDING REMARKS

The examples chosen for this paper have illustrated the ability of the modelling system to predict wind fields in complex terrain and to give a qualitative estimate of the dispersion of pollutants by those winds. Such a system is ideal as an aid in selecting locations for emitters in regions of little or no data. As far as established sources are concerned, the system can be used to identify those meteorological conditions under which air quality can be improved by a short-term reduction in emissions. Alternatively, the effect of an increase in the number of emitters in an area can also be evaluated.

A more quantitative approach to predicting the concentration fields involves dividing the model domain into boxes and counting the particles in each box. However, care must be taken with both the space and time averaging necessary in this approach and the number of particles that must be released before representative results can be obtained. An alternative scheme to the counting of particles approach is the kernel density estimation technique which offers more sophistication and a decrease in the number of particles needed to obtain a given accuracy in concentration estimates.

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1 Introduction

Puff modeling is a procedure for determining the concentration of a pollutant plume in a given area. It is a numerical model that simulates the physical processes of advection, diffusion, and deposition. The model is based on the assumption that the plume is a continuous body of fluid that moves and spreads as it travels downwind. The model is used to predict the concentration of a pollutant at a given location and time. The model is based on the following assumptions: (1) the plume is a continuous body of fluid; (2) the plume moves and spreads as it travels downwind; (3) the concentration of the pollutant is uniform throughout the plume; (4) the concentration of the pollutant is zero outside the plume; (5) the concentration of the pollutant is constant in time; (6) the concentration of the pollutant is constant in space; (7) the concentration of the pollutant is constant in time and space; (8) the concentration of the pollutant is constant in time and space; (9) the concentration of the pollutant is constant in time and space; (10) the concentration of the pollutant is constant in time and space.

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THESE ARE THE ASSUMPTIONS OF THE PUFF MODEL. IT IS A SIMPLE MODEL THAT CAN BE USED TO PREDICT THE CONCENTRATION OF A POLLUTANT AT A GIVEN LOCATION AND TIME. THE MODEL IS BASED ON THE FOLLOWING ASSUMPTIONS: (1) THE PUFF IS A CONTINUOUS BODY OF FLUID; (2) THE PUFF MOVES AND SPREADS AS IT TRAVELS DOWNWIND; (3) THE CONCENTRATION OF THE POLLUTANT IS UNIFORM THROUGHOUT THE PUFF; (4) THE CONCENTRATION OF THE POLLUTANT IS ZERO OUTSIDE THE PUFF; (5) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN TIME; (6) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN SPACE; (7) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN TIME AND SPACE; (8) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN TIME AND SPACE; (9) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN TIME AND SPACE; (10) THE CONCENTRATION OF THE POLLUTANT IS CONSTANT IN TIME AND SPACE.

They are:

1. Simulation of the physical processes (advection, diffusion, and deposition).
2. The puff model is a simple model that can be used to predict the concentration of a pollutant at a given location and time.

Modelling of Flow and dispersion in a Coastal Area

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1 Introduction

Puff modelling as a principle has for long time been recognized as a fast and convenient method for calculating atmospheric dispersion in connection with risk and safety assessment studies for non-homogeneous, but flat terrain, including non-stationary dispersion scenarios. Individual puff growth rates can here conveniently be assigned according to the local level of turbulence, and advection of the cloud can here be interpolated directly from measurements of the mean wind speed and direction at representative locations in the terrain (see e.g. Mikkelsen et al. 1984, Thykier-Nielsen and Mikkelsen 1987).

Turning to the case of orographically influenced regions, including the case of complex terrain, observations of smoke diffusion often exhibits local scale plume bifurcation and layer decoupling, including significant channeling and local valley flows. Additional information is here required about the dispersing wind field, achievable for instance; by coupling the puff-model to a high-resolution mean flow model. This provides an opportunity for introducing a scale separation in the dispersion calculation, ideally, a high-resolution grid flow model provides a three-dimensional mean flow field in response to the underlying terrain, which in the dispersion model can be used directly for advection. Then, the dispersion process of the clouds is conveniently calculated on a subgrid scale, at least as long as the size of the cloud is smaller than the grid spacing. On the sub-grid scale, it is then assumed that the dispersion process is locally homogeneous, for which case the growth rate to a good approximation can be parameterized in terms of the local level of turbulence (see e.g. Mikkelsen et al., 1987).

RIMPUFF-Risø Mesoscale Puff model is able to handle flat as well as complex terrain. In this paper two examples have been chosen to illustrate the use of the model in coastal areas.

They are:

1. Simulation of the Øresund experiment (flat, inhomogeneous terrain).
2. The Vandenberg study (complex terrain).

2 Risø Puff Diffusion Model

The Risø puff diffusion model is operational as a computer code used for the prediction and/or simulation of the diffusion and advection of atmospheric pollutants. The puff model simulates a plume by releasing Gaussian shaped puffs at a fixed release rate on a specified grid. The amount of material in a single puff equals the release rate times the elapsed time between puff releases.

The location of the puffs on the grid is determined by computing their movement for finite time steps using the measured wind field based on a network of wind measurement stations. The growth of the individual puffs is computed from simultaneous specifications of the atmospheric turbulence intensity and stability in the dispersion area. The height of the inversion cap through which pollutants cannot pass and the source height can easily be adjusted. Grid distances within the model may vary from meters to kilometers, and time durations from seconds to hours are possible.

The puff model has the capability of monitoring multiple sources of puffs, and its grid may contain hundreds of puffs. A puff source can be located anywhere on the grid and can have an individual release rate, release time and buoyancy production.

For each advection step, the model calculates the concentration at each grid point by summing the contributions from surrounding puffs. The grid concentrations can be allowed to accumulate or simply be updated with the latest instantaneous value.

The output of the model yields the puff-locations and concentrations at time intervals controlled by the input data.

The disc-stored output data are further evaluated by a graphic program which creates background maps, wind-field plots, puff-plots and iso-concentration contours.

The puff model code is available in FORTRAN77. Further information on the Risø diffusion model is available in (Mikkelsen, 1982; Mikkelsen et al., 1984; Thykier-Nielsen et al., 1985 and Mikkelsen et al., 1987).

3 Simulation of the 1984-Øresund Experiments

3.1 Introduction

RIMPUFF shares with most other mesoscale models that its experimental evaluation has so far been restricted to data representative only of short to medium range diffusion conditions. Because mesoscale dispersion is controlled by features and parameters often to be neglected on the small or local scale, simple extrapolations into the mesoscale range cannot readily be justified, except under very exceptional conditions. A thorough experimental evaluation of the mesoscale calculations are therefore essential in order to establish the models credibility also on this scale. This is an important step before more general environmental assessment studies can be based on mesoscale model simulations.

We present simulations of 3 of the 1984 Øresund tracer experiments (Gryning, 1985), which took place over an inhomogeneous region (land/water/land) at distances typical for the mesoscale. The experiments were generally carried out during strong and steady winds except for the one on 14 June when the winds were quite variable due to a steady counter clockwise rotation of the wind during the experiment.

3.2 Modelling

3.2.1 Wind Field Calculation

In order to calculate the mesoscale wind field of a region based on a network of available observations, we use the method of objective wind analysis, i.e. weighted interpolation on a regular grid. A $1/r^2$ weighting function was used where r is the distance from the grid point to a measurement station.

3.2.2 Diffusion Parameters

The diffusion of a puff has been found to significantly depend upon the turbulence intensity. The puff model is based on the principle of relative diffusion, as for instance described in (Pasquill et al., 1983). The turbulence intensity is introduced in the calculations in the form of standard deviations of the lateral and vertical wind direction fluctuations, (or alternatively, inferred from the Pasquill-Turner scheme).

The turbulence intensities in the lateral and vertical directions are to a close approximation equal to the standard deviation of the wind direction fluctuations in the lateral, σ_θ , and the vertical, σ_ϕ , directions. The standard deviations are averaged over the same time periods as the wind speed measurements used for advecting the puffs (typically 10 min). Therefore, intensities for the turbulence, are updated with the advecting wind speed vector after each time step. The area of simulation defined in a grid, can contain different stability regions. The growth rate of a puff will therefore reflect changes in atmospheric stability depending on its position in the grid.

3.3 Puff Simulation of the Experiments

Nine tracer experiments were carried out during the Øresund-experiment; however, only six of these were judged to be suitable for this model evaluation study. Of the three remaining experiments, two were insufficiently covered by the tracer sampling arcs, while the tracer studies were cancelled during the third experiment due to unfavorable weather conditions.

Stability Areas

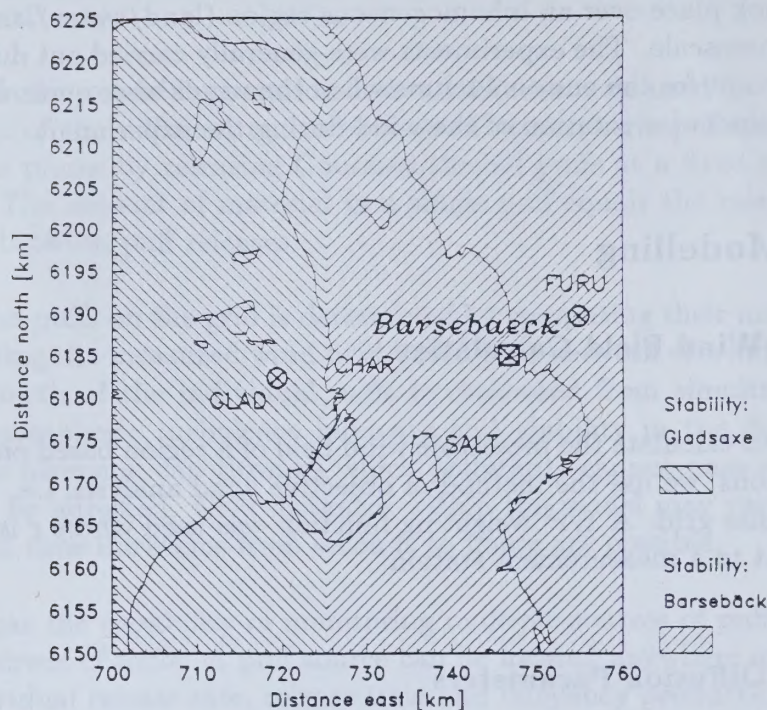


Figure 1: The stability area chosen for the calculations with RIMPUFF. The positions of the 3 meteorological stations Gladsaxe (GLAD), Barsebäck, and Furulund (FURU) are also shown. Coordinates in UTM zone 32 are used.

3.3.1 Input Data Conditioning

Meteorological input parameters to RIMPUFF consisted of 1) the wind speed and direction at two positions in the area; and 2) the standard deviation of the wind fluctuations in both the lateral and vertical directions.

In the simulation of the Øresund tracer experiments we used 10 min. averaged meteorological measurements performed at Barsebäck or Furulund and Gladsaxe, (see Fig. 1).

At Barsebäck the 10 min. averaged measurements of wind speed and direction at 95 m height were used. The 10 min. averaged standard deviation of the lateral and vertical wind fluctuations were derived from the turbulence measurements at the same height. At Gladsaxe the turbulence measurements at 115 m height were used to derive 10 min. averages of wind speed, direction and variances of lateral and vertical wind fluctuation.

3.3.2 Simulations with the Puff Diffusion Model

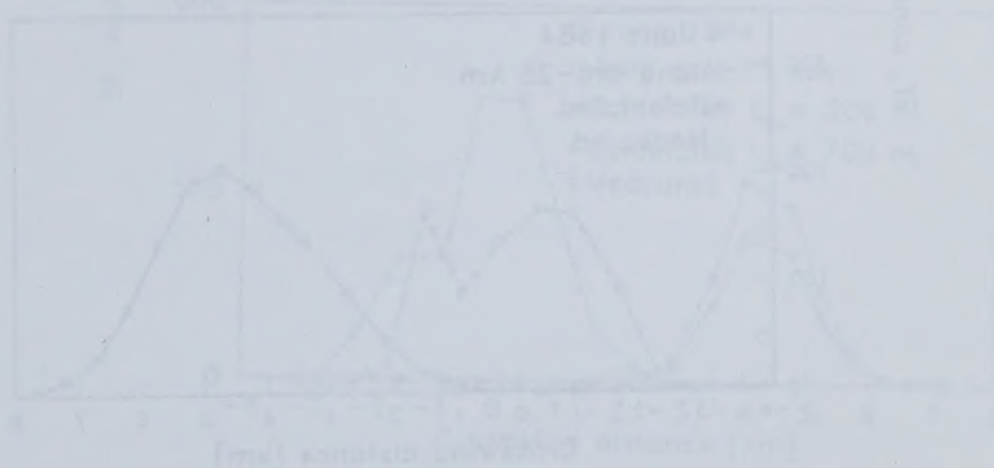
In simulating the tracer experiments it was not straight-forward how the complicated structure of the atmospheric stability over Øresund should be presented to the model. We have chosen to let the dispersion over the entire Øresund be controlled by the conditions at Barsebäck. Inland from the coast of Sjælland, dispersion characteristics

were controlled by the conditions at Gladsaxe. This is illustrated in Fig. 1. The procedure does not take into account the stable layer close to the water surface over Øresund that inhibits dispersion and affects the wind field in the lower part of the atmosphere (Doran et al., 1987). However, above this stable layer the air is neutral or unstable and corresponds to upwind land conditions, although the turbulence is weaker because it is decaying. Fig. 2-4 shows the results of this type of simulation of the 3 of the tracer experiments. The figure also shows some of the positions at which the 1 hour averaged tracer concentration measurements were carried out.

In Fig. 2-4 we have compared the results of 3 of the simulations with the actually measured tracer concentrations. In all the experiments there is good agreement between the position of the measured and the simulated plumes. This shows that the wind field is well predicted by the model. It can also be seen from Fig. 2-4 that both the height and the width of the simulated plume is in reasonable agreement with the measurements in all experiments.

The sensitivity of the model simulations to the height of the mixing layer is illustrated in Fig. 3. Decreasing the height of the mixing layer from 700 m to 200 m decreases the ratio between the measured and the predicted peak values of the concentration from a factor of 3 to 2.

When comparing the simulated and actually measured concentrations it is characteristic that the agreement, irrespective that the stable internal boundary layer over Øresund is not taken into account in the modelling procedure, is good, considering the extremely complicated nature of the meteorology in the area. It is therefore surprising that the puff model, even under such complicated meteorological situations produces rather accurate estimates of the concentration field.



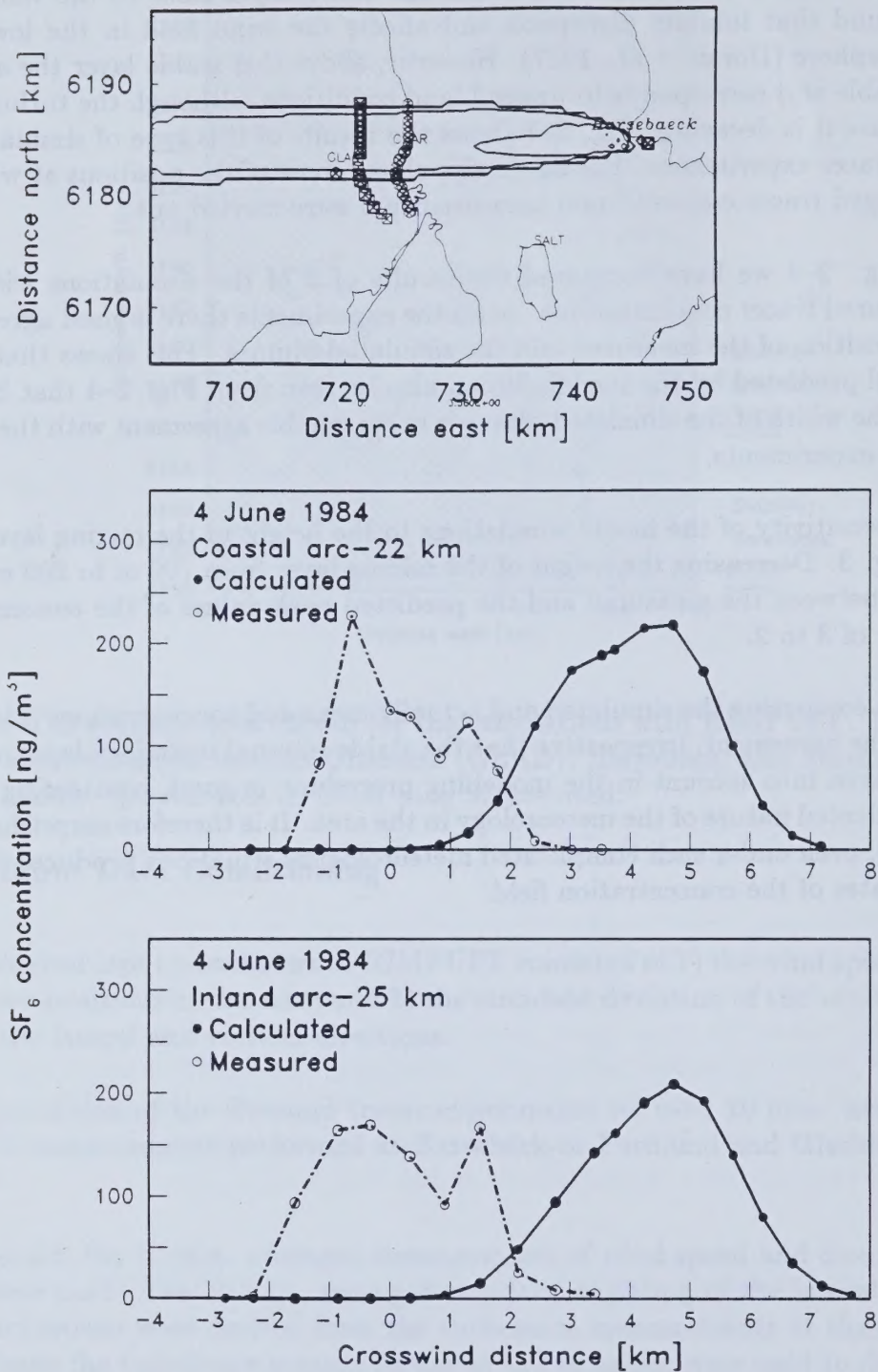


Figure 2: The calculated plume concentration pattern with the position of the sampling arcs overlaid (top figure), and a comparison of the calculated and measured concentrations along the coastal arc (middle figure) and the inland sampling arc (bottom figure). Coordinates on the top figure is in UTM zone 32.

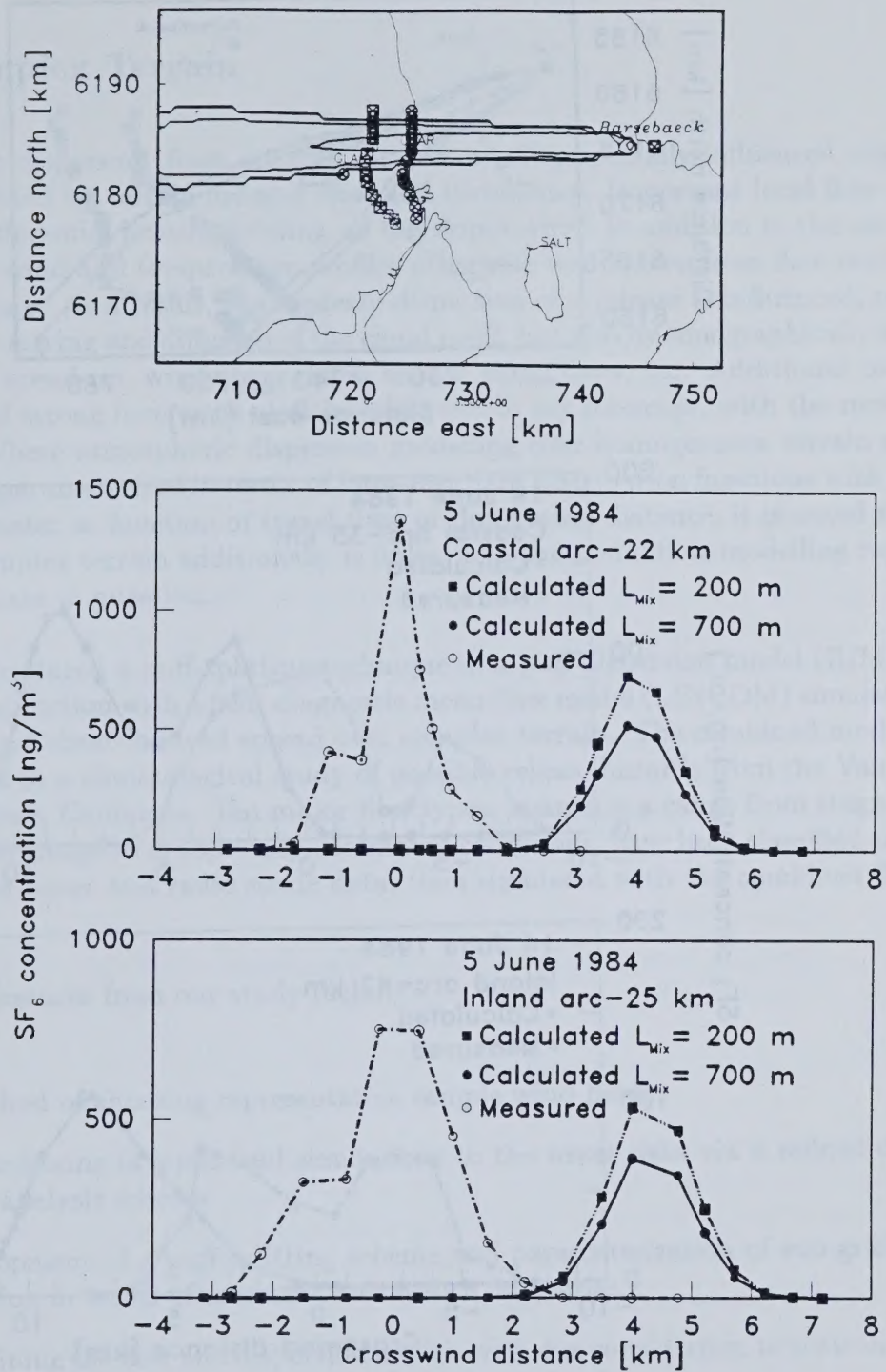


Figure 3: The calculated plume concentration pattern with the position of the sampling arcs overlaid (top figure), and a comparison of the calculated and measured concentrations along the coastal arc (middle figure) and the inland sampling arc (bottom figure). Coordinates on the top figure are in UTM zone 32.

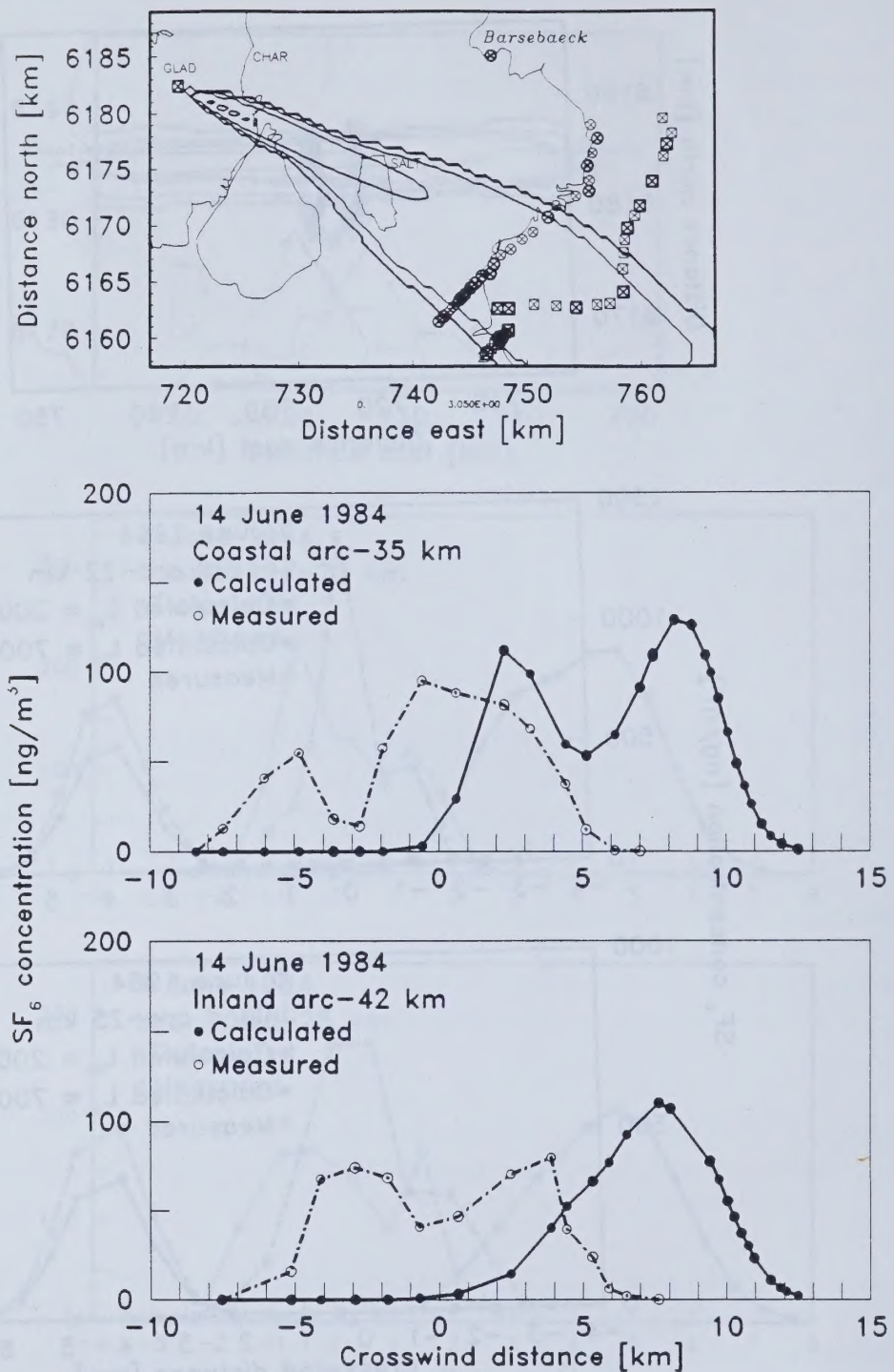


Figure 4: The calculated plume concentration pattern with the position of the sampling arcs overlaid (top figure), and a comparison of the calculated and measured concentrations along the coastal arc (middle figure) and the inland sampling arc (bottom figure). Coordinates on the top figure is in UTM zone 32.

4 The Vandenberg Study

4.1 Complex Terrain

Atmospheric dispersion from sources located in orographically-influenced regions is heavily modified by terrain-induced flow and turbulence. Important local flow-driving forces are differential heating/cooling on the slopes which in addition to the orography disturbs the wind and temperature profiles otherwise well-known from flow over homogeneous terrain. As a result, atmospheric dispersion of a release is influenced, not only by turbulent mixing and diffusion of the cloud itself, but also by topographically-induced channelling, speed-up, wind shear, local terrain slope flows, etc. Additional complications occur if strong inversions aloft interfere not to say intercept, with the mountains, hills, etc. Where atmospheric dispersion modelling over homogeneous terrain successfully can be parameterized in terms of form-invariant distribution functions with a single sigma-parameter as function of travel time or down-wind distance, it is noted that diffusion in complex terrain additionally is linked to a successful flow-modelling capability over the terrain in question.

We have introduced a puff-splitting technique in a puff dispersion model (RIMPUFF), which in conjunction with a fast, diagnostic mean-flow model (LINCOM) simulates both bifurcated and shear-induced spread over complex terrain. The combined models have been applied in a climatological study of possible release hazards from the Vandenberg Air Force Base, California. Ten major flow types, spanning a range from stagnation to sea breeze, drainage to synoptically dominated conditions, have been classified and characterized via tower and radio sonde data, then simulated with the combined flow/puff models.

Particular features from our study include:

1. a method of choosing representative sample wind fields
2. the anchoring of wind-field simulations to the tower data via a refined objective wind analysis scheme.
3. development of a puff-splitting scheme and parameterization of sub-grid relative diffusion in terms of measured turbulence intensities.
4. combining the flow and dispersion models with due consideration to scale-dependent parameterization.

4.2 The Climatology Study

Vandenberg AFB lies in the California coastal transition zone in complex terrain bounded by the Santa Ynez mountains (Fig. 5). Its complex meteorology has been studied for some time through computer and physical model simulations, field studies, and an on-line array of meteorological sensors.

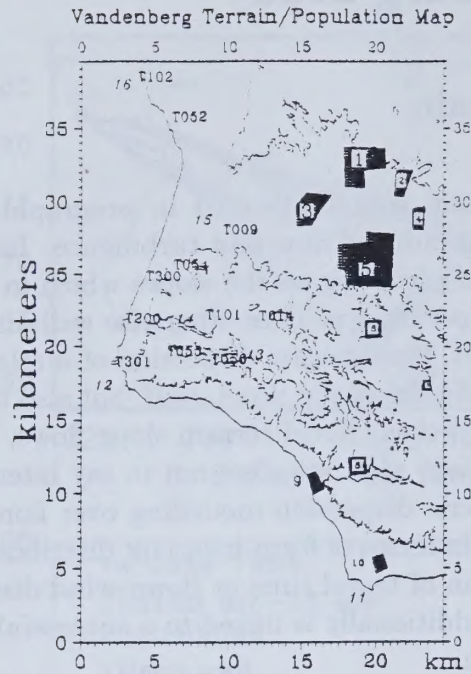


Figure 5: Vandenberg terrain/population map.

- | | |
|-------------------------|-----------------------|
| 1. Vandenberg village | 9. Jalama Ranch |
| 2. Mission Hills | 10. Cojo Ranch |
| 3. Lompoc Prison | 11. Point Conception |
| 4. La Purisma Mission | 12. Point Arguello |
| 5. Lompic | 13. Tranquillon Ridge |
| 6. White Hills Hospital | 14. Honda Canyon |
| 7. Jalama School | 15. Santa Ynez Valley |
| 8. Jalama Beach | 16. Point Sal |

T102, etc mark the wind tower locations.

To determine characteristic flowtypes and their forcing 7200 hourly averaged flow pattern were studied. From these patterns we designed wind direction templates at the 44-foot level for 10 flow types including stagnation, land/sea breeze, drainage and synoptically dominated conditions. These templates were used to classify all hourly averaged fields, see Fig. 6.

4.3 Modelling

4.3.1 Flow Model

With operational risk assessment in mind, we adopted the extremely fast diagnostic non-hydrostatic flow model LINCOM (Troen and de Baas, 1986), in which linearization of the equations of motion leads to a system conveniently solved using a spectral decomposition of the horizontal scales involved. For a linearized approach to be a good

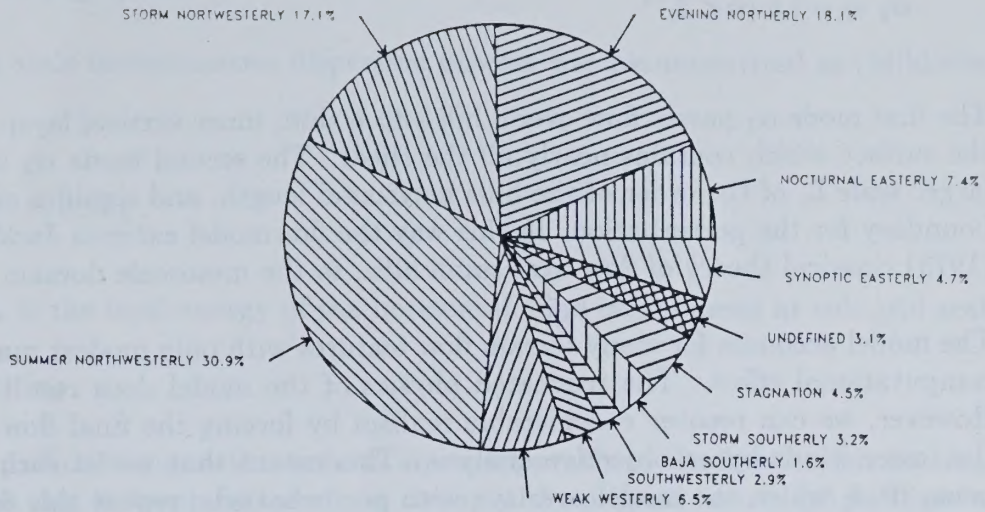


Figure 6: Hourly flow type percentage distribution from 9/83-8/84

approximation, it is required that the aspect ratio of the terrain (typical height/typical length) must be small compared to unity. Even then, the terrain features will create significant perturbations of pressure and local stress which in turn will cause significant perturbations in the wind velocity near the surface.

For a homogeneous boundary layer with no buoyancy effects, the vertical profile of the mean wind U will be logarithmic, and the turbulent diffusivity for momentum will be linear, viz.: $K(z) = \kappa u_* z$. Hence κ is the von Karman constant and u_* is the surface friction velocity.

In LINCOM the vertical variation of U and K are resolved for topography by assuming a spectral dependance

$$U_o = U_o(c_1 | \alpha |^{-1}) \quad (1)$$

$$K = K(c_2 | \alpha |^{-1})$$

where α is the vertical wavenumber. Thus, each "mode" of the velocity and diffusivity associates with a vertical length scale of order $\sim | \alpha |^{-1}$. This means that the long modes generated by larger scale terrain features will reach greater heights (small $| \alpha |$) and will "feel" the larger turbulent diffusivities and wind velocities found at these heights, while smaller diffusivities and velocities drive the less penetrating modes.

For a particular horizontal wavenumber k , the solution for the perturbation velocity can be shown to depend on two different values of the vertical wave numbers α_1 and α_2 :

$$\alpha_1 = \frac{\sqrt{2}}{2}(i-1)\ell^{-1} \text{ with } \ln\left(\frac{c_1\ell}{z_o}\right) = c_2 k^2 L \quad (2)$$

and

$$\alpha_2 = - |k| < L^{-1}$$

The first mode α_1 parses flow over a hill into a thin, inner vertical layer of depth ℓ near the surface which contains nearly all the shear. The second mode α_2 involves a much larger scale L , of the order of the hills horizontal length, and signifies an outer scale or boundary for the perturbation. In this way the flowmodel extends Jackson and Hunt's (1975) classical theory of flow over small hills, to the mesoscale domain.

The model accounts for many terrain flow features with only modest mathematical and computational effort. The truncated physics of the model does result in some error. However, we can recover considerable realism by forcing the final flow field to match the tower winds by an objective analysis. This means that we let each tower define a mean field, which the model overlays with perturbations, repeat this for all n towers, then average the n perturbed fields to get a final field. In the averaging, the averaging of the n field falls by the weighted square of the distance from the tower upon which the field is based. The weighting warps the circle of influence from each tower into ovals whose aspect ratio and size differ with terrain.

4.3.2 Dispersion Model

The mesoscale dispersion model RIMPUFF is a standard gaussian puff model equipped with computer-time effective features for terrain and stability-dependent dispersion parameterization, plume rise formulas, inversion and ground-level reflection capabilities and wet/dry (source) depletion, (Mikkelsen et al. 1984). In addition, the code optionally provides local relative diffusion parameterization and a scheme for horizontal/vertical shear diffusion.

For complex terrain (i.e. for combination with LINCOM) RIMPUFF was additionally provided with a puff-splitting algorithm, in order to simulate bifurcated diffusion and topography-induced plume shearing.

Horizontal puff spread

Sub-grid scale instantaneous dispersion process was parameterized as (Mikkelsen et al. 1987).

$$\frac{d\sigma_p}{dt} = 0.3\sigma_h \quad (3)$$

where σ_h is the local energy of the horizontal wind components at sub-grid scales.

Puff-splitting

When a puff during growth reaches a size comparable to the flow models horizontal grid resolution Δg , it “pentafurcates”, i.e. we split the original (gaussian) puff into a cluster of five new, but smaller (gaussian) puffs, according to:

$$\begin{aligned} \text{i)} \quad & \sum_5^2 = \sigma_{p1}^2 \\ \text{ii)} \quad & C_5(0,0) = C_1(0,0) \\ \text{iii)} \quad & \sigma_{p5} = \frac{1}{2}\sigma_{p1} \\ \text{iiii)} \quad & \text{Mass conservation} \end{aligned} \quad (4)$$

That is:

i) The second moments (inertia moment about the centroid) before (σ_{p1}) and after ($\sum_5^2$) pentafurcation are equal. ii) The center (peak) concentration are equal. iii) The size of the pentafurcated puffs are reduced to one half. iv) the total mass of pollutant is conserved.

Fig. 7 shows the pentafurcation principle: Five new, one central and four satellite puffs, replaces the original single puff of size σ_{p1} . Four satellite puffs are dislocated the distance $0.89\sigma_{p1}$ away from the origo, each carrying 23.53% of the total mass. A fifth, assigned the remaining 5.8% of the mass is still located at the origin.

A similar, but one-dimensional puff-splitting scheme fulfil the constrains in Eq. 4 for the vertical direction. Here, only three new puffs are required (trifurcation) two satellite puffs located at $\pm 1.9\sigma_1$, each containing 26.58% mass, and one center-located, containing the remaining 48.84% mass.

4.4 Model Simulations

Meander is difficult to predict. However, for releases lasting less than two hours, analysis shows that the meander usually remains with a given flow type. Thus, we illustrate this by showing the plausible range of directions and transport distances for each flow type. To estimate such ranges, one hundred hourly averaged wind fields (ten per flow type) were chosen from a one-year set of 7200. For a representative sampling, all n -fields were ranked from 1 to n within a given flow type by wind angles at the most consistent, least terrain influenced tower (T052). Then, for the simulations we chose 10, having the ranks; $n/20, 3n/20, 5n/20 \dots 19n/20$.

Penta Puff

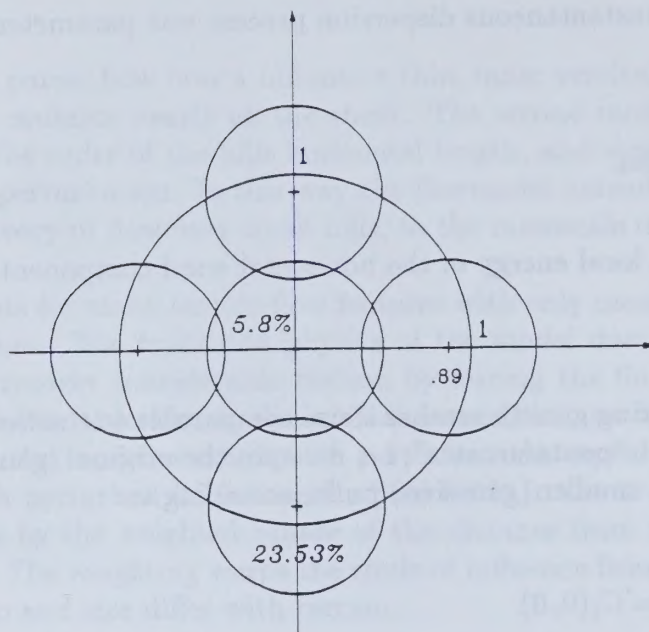


Figure 7: Principle in horizontal puff-splitting scheme

Figures 8–10 are samples from 300 such simulations. Apart from the population/terrain map shown in Fig. 5 for a given flow type and release site, we also show a) 10 center-line trajectories based on the 10 wind fields selected above; b) the extreme backed wind field; c) its matching dose isopleth map; d) the extreme veered wind field; and e) its matching dose isopleths. the backed, middle, and veered wind fields correspond to those having the ranks: $n/20$, $9n/20$, and $19n/20$. The set of trajectories conveys the plausible angular range of meander within a given flow type (each driven by a separate wind field), and each isopleth map simulates a single pollutant release footprint. However, the pollutant toxic hazard corridor (THC) includes meander. Thus, its spread is defined by the critical isopleths on the veered and backed maps. Its downwind range is given by a compass arc, centered at the release point, which grazes the downwind edge of the furthest critical isopleth (normalized by current wind speed). The critical isopleth values depend on the release situation. They change for different chemicals, release rates, exposure limits, and evaporation times.

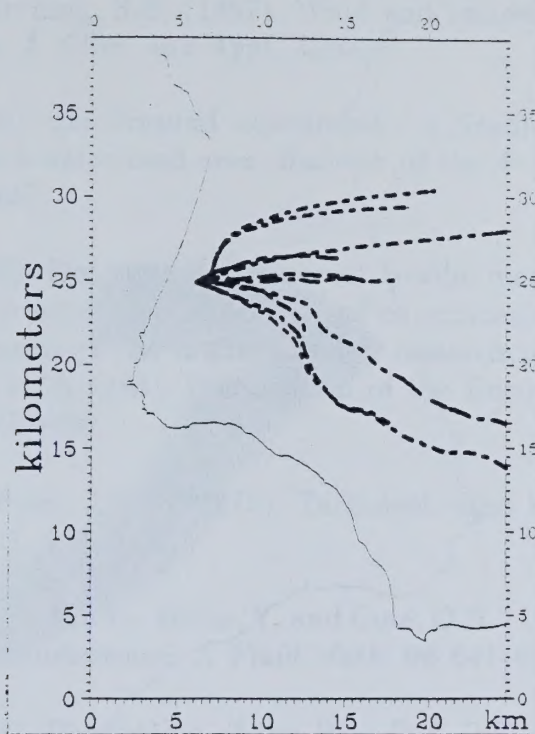


Figure 8: Trajectories for sea breeze westerlies from near tower 300.

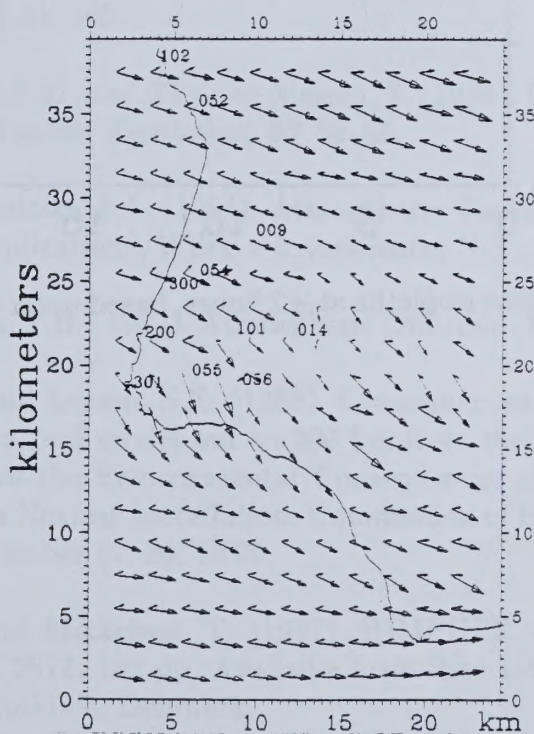


Figure 9: Sample extreme veered and backed, westerly sea breeze winds.

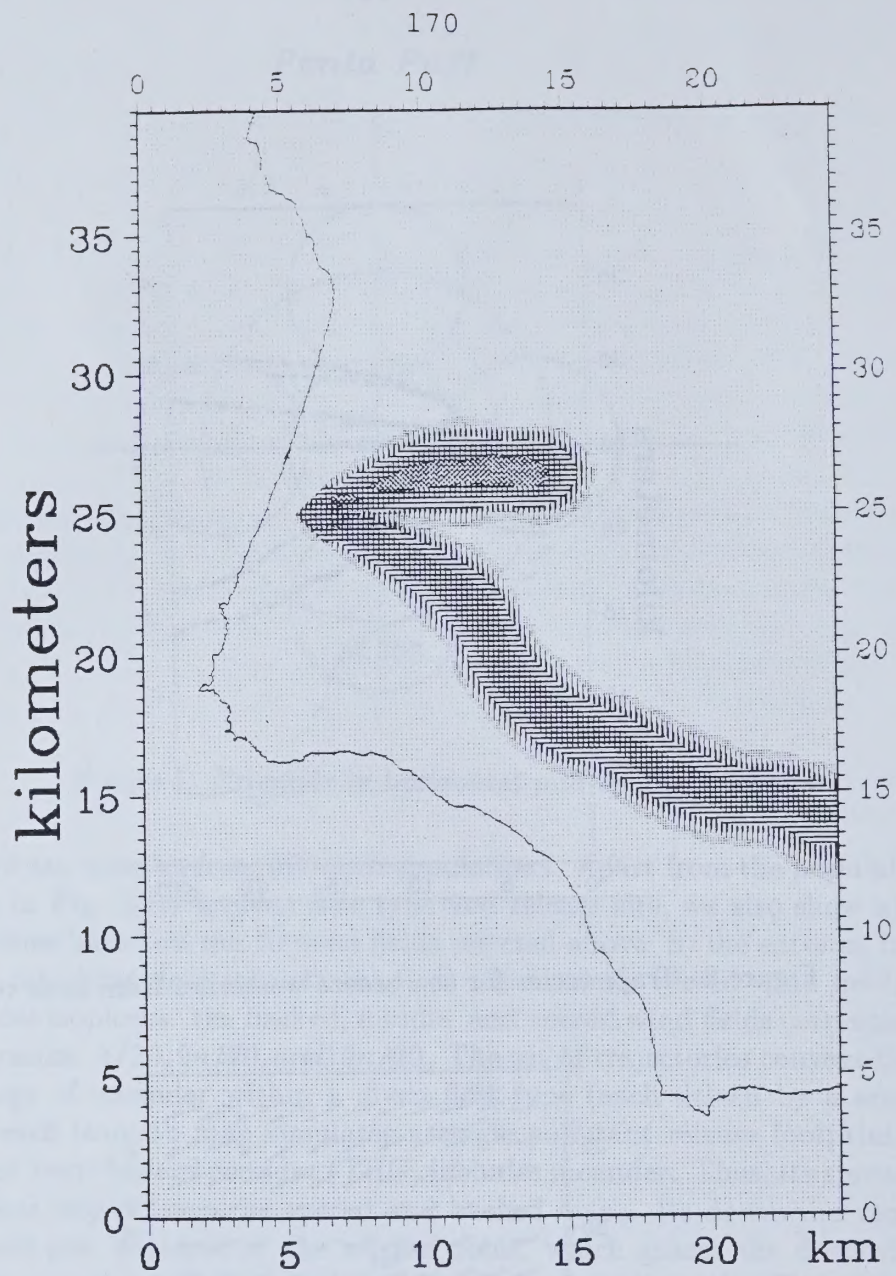


Figure 10: Sample dose isopleths at +2 hours, based upon veered and backed wind fields from Fig.9

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MODELING THE LOCAL ATMOSPHERIC CIRCULATION AND POLLUTION DISPERSION IN THE VICINITY OF THE ZARNOWIEC NUCLEAR POWER PLANT

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1. INTRODUCTION

In this decade, with the Bhopal and Chernobyl accidents, meteorology has come to play an increasing role in emergency response planning, particularly for those facilities handling toxic chemicals and radiological materials. The majority of emergency response modeling systems rely on Gaussian dispersion models based on the assumptions of a stationary and homogeneous atmosphere. However, such an approach has serious shortcomings in coastal zones, especially within 15 km of the shoreline, where wind field and turbulence exhibit large spatial variability. A significant improvement may be achieved by application of a synoptic classification scheme and a mesoscale meteorological model coupled with a Lagrangian particle dispersion model (Pielke et al., 1987). Lyons and Fisher (1988) identify several coastal mesoscale regimes with different transport and diffusion conditions. They propose a conceptual framework for allocating resources in developing an adequate emergency response system which should include three major factors: (1) the frequency of mesoscale regimes; (2) the extent to which the regime can result in high concentration; (3) the ease with which it can be modeled with due considerations given for required input data. The frequently occurring events with the potential for significant impacts, and which are relatively easy to model should be given preference. This methodology allows one to focus efforts with given resources (time and funds) as well the limit of the current state-of-the-art of meteorology.

The above ideas will be followed in developing the emergency response modeling system for the nuclear power plant under construction on the shore of Lake Zarnowiec in northern Poland

(Fig.2). Lake Zarnowiec lies in the coastal zone of the Baltic Sea in a valley oriented to SE in direction of large cities. The area of interest is characterized by influence of sea and land breeze circulation, hilly terrain (height differences up to 150m), and strong nonhomogeneities of the land surface.

The meteorological research program for this area consists of field experiments and theoretical studies using a mesoscale meteorological model. Meteorological measurement facilities available at Zarnowiec include a standard meteorological station, 200m meteorological tower, a radar for tracking tetroons and a standard radiosounding station at Łeba 30 km westward from the power plant. Several measurement campaigns have been carried out during recent years using tethered and pilot balloons. A series of tracer dispersion experiments with SF₆ for different spatial scales are planned for the next three years (1989-1991). Simultaneously, a monitoring network of several stations with radioactivity and meteorological measurements is under planning and construction.

Numerical mesoscale modeling is expected: (1) to support planning dispersion field experiments and monitoring network, and (2) to provide recommendations for development of an emergency dispersion model to be used in the power plant. The paper presents a current state of developing the mesoscale dispersion modeling system and its application for the Zarnowiec area. A preliminary evaluation of the mesoscale model was performed using data from the Öresund experiment and is described in a companion paper (Uliasz, 1988a).

2. MESOSCALE METEOROLOGICAL MODELING SYSTEM

The system of numerical mesoscale models (Fig.1) designed for the use on personal computers is under development at Department of Meteorology, Technical University of Warsaw. The objective of the system is investigating atmospheric flow and air pollution dispersion in the γ -mesoscale including effects of terrain topography, large water bodies and urban areas.

The main part of the modeling system is the three-dimensional γ -mesoscale meteorological model with the following characteristics and configuration:

- primitive, hydrostatic equations formulated for mesoscale perturbations, 2- or 3-dimensional;
- terrain following coordinates;
- dry thermodynamics (no liquid water, vapor only);
- turbulence parameterization: second-order closure - level 2.5 scheme proposed by Mellor and Yamada (1982) and modified for the case of growing turbulence by Helfand and Labraga (1988);
- surface temperature: (1) prescribed or (2) calculated from the force-restore equation with surface radiation flux parameterization adopted from Van Ulden and Holtslag (1985);
- initialization with 1-dimensional model version;
- numerical algorithm: finite difference implicate scheme, two-cycle splitting method;
- vertical resolution: 15 levels up to 2 km;
- horizontal resolution: telescoping grid mesh with $\Delta x = 2\text{km}$ in the centre of modeling domain;
- time step of integration: $\Delta t = 120\text{s}$
- computation time for 1 hour of simulation and memory requirements (IBM PC/AT, 10MHz):
 - (1) $15 \times 12 \times 15$ gridpoints, 0.64 hour, 237 KB,
 - (2) $27 \times 23 \times 15$ gridpoints, 2.87 hour, 547 kB.

The mesoscale model creates wind and turbulence fields required as input data by dispersion models. The 1-dimensional model version can be applied for a flat and homogeneous terrain.

The Lagrangian dispersion model included in the modeling system allows one to simulate releases of a nonbuoyant and passive pollution from single point sources by tracking the motion of particles (Monte-Carlo model) or Gaussian puffs. The particle version of the dispersion model can be used as a visualization technique for the simulated atmospheric flow in complex terrain.

The models are associated by a program environment containing programs for data preprocessing and a graphical presentation of results. All programs are written in the FORTRAN-77 standard and implemented on the IBM PC/AT compatible micro-computer. The modeling system is used not only for scientific purposes but also as a demonstration tool for university students. The detailed description of the modeling system is available from

the author (Ullasz, 1988b).

3. SIMULATIONS FOR THE ZARNOWIEC AREA

The most complicated pattern of air pollution dispersion in the Zarnowiec area can be expected under weak wind conditions with strong thermal contrast between sea and land. Pollution dispersion is driven then by an interaction of synoptic flow and sea/land breeze circulation. As an example the results of numerical simulation performed for a cloudless summer day with a weak western geostrophic wind (5m/s) are presented below. The simulation was carried out from 0 to 14:00 CET with $15 \times 12 \times 15$ grid points and $\Delta x = 2.5\text{km}$ in the centre of considered area.

Fig.3 shows simulated surface wind fields on morning and on afternoon. Some disturbances of wind field at 10:00 are due to terrain topography and surface nonhomogeneity. Four hours later when the land surface becomes about 10 degrees warmer than the sea a visible effect of sea breeze appears with the onshore wind component up to 15km inland.

The simulated fields of mean wind and velocity variances were used to run the particle version of the dispersion model for continuous pollution releases at Zarnowiec from the heights 50m and 300m. Particles were released starting at 6:00 with the emission ratio 30 particles/hour for centerline calculations and 240 particle/hour for plume simulation.

The centerlines of pollution plumes are obtained by tracking a set of particles without taking into account the turbulence. Fig.4 illustrates significant changes of the plume direction during the day. The rise and descending of the nonbuoyant plume are caused by upward motions over warm land and downward motions over cold sea respectively. The plumes simulated including turbulence (Fig. 5 and 6) show a significant lateral dispersion due to spatial and temporal variability of wind. Fig. 7 presents the plume centerlines calculated the 3-dimensional meteorological fields (as in Fig.4) and the centerlines calculated using the 1-dimensional meteorological profiles taken at Zarnowiec. This comparison clearly demonstrates that any dispersion model based on the on-site meteorological measurements can not correctly predict

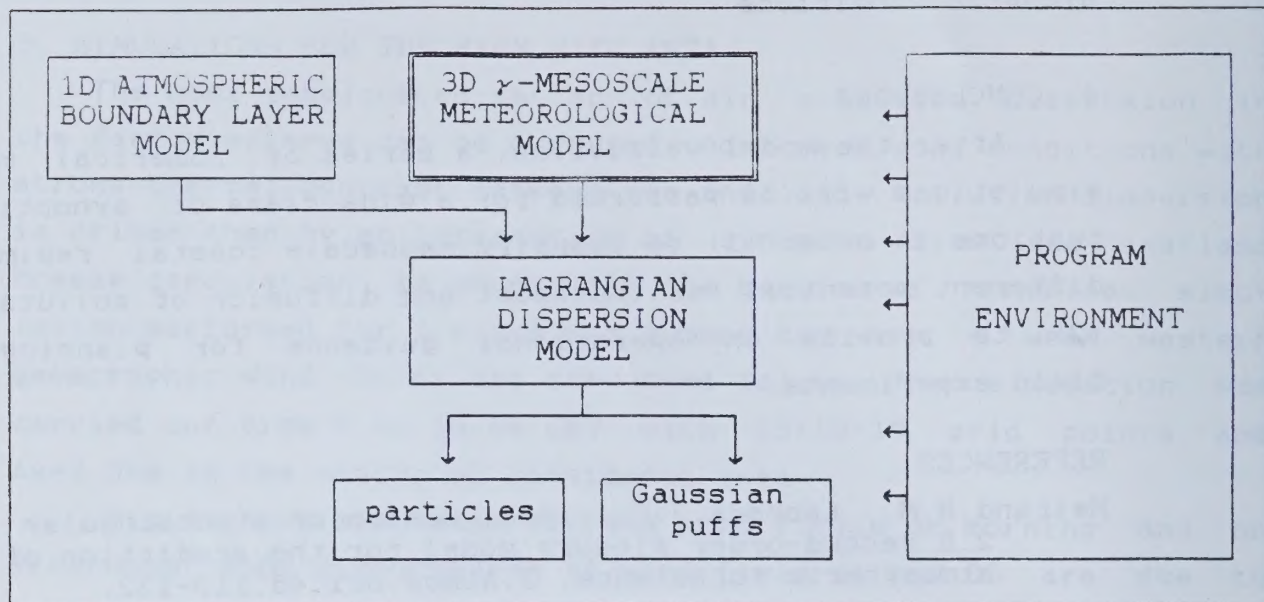
plume or puff transport in the coastal area under considered synoptic conditions.

4. CONCLUSIONS

After the model validation, a series of numerical mesoscale simulations will be performed for a wide class of synoptical situations in order (1) to identify mesoscale coastal regimes with different potential for transport and diffusion of pollutants; and (2) to provide an operational guidance for planning tracer field experiments.

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INPUT

land/water boundaries
 terrain height
 land usage
 water surface temperature
 geostrophic wind
 radiosounding
 cloud cover
 day of year

point pollution source:
 coordinates
 emission ratio

OUTPUT (3-dimensional time dependent fields)

wind u, v, w
 potential temperature
 specific humidity
 turbulent energy
 wind velocity variances
 eddy diffusivities

pollution concentration
 positions of particles

FORMATS

XY plane sections
 XZ, YZ vertical sections
 X, Y, Z profiles

Fig.1. Mesoscale dispersion modeling system

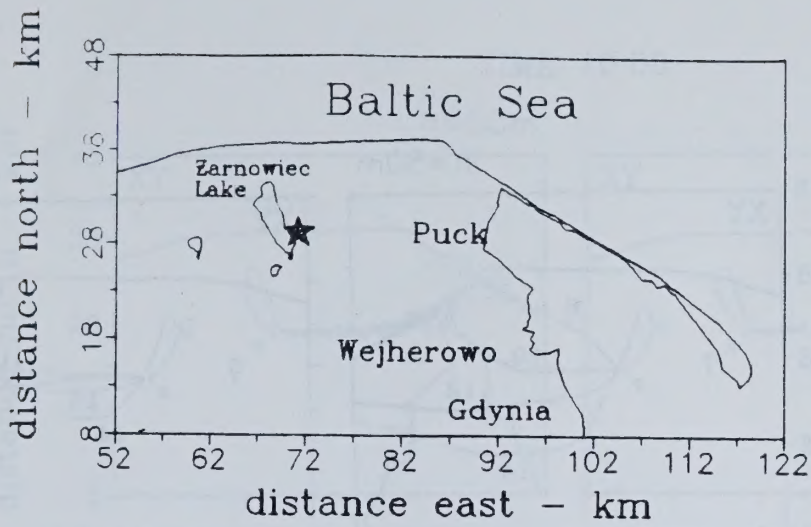


Fig.2. Coastal zone of the Baltic Sea in vicinity of the Zarnowiec nuclear power plant (*)

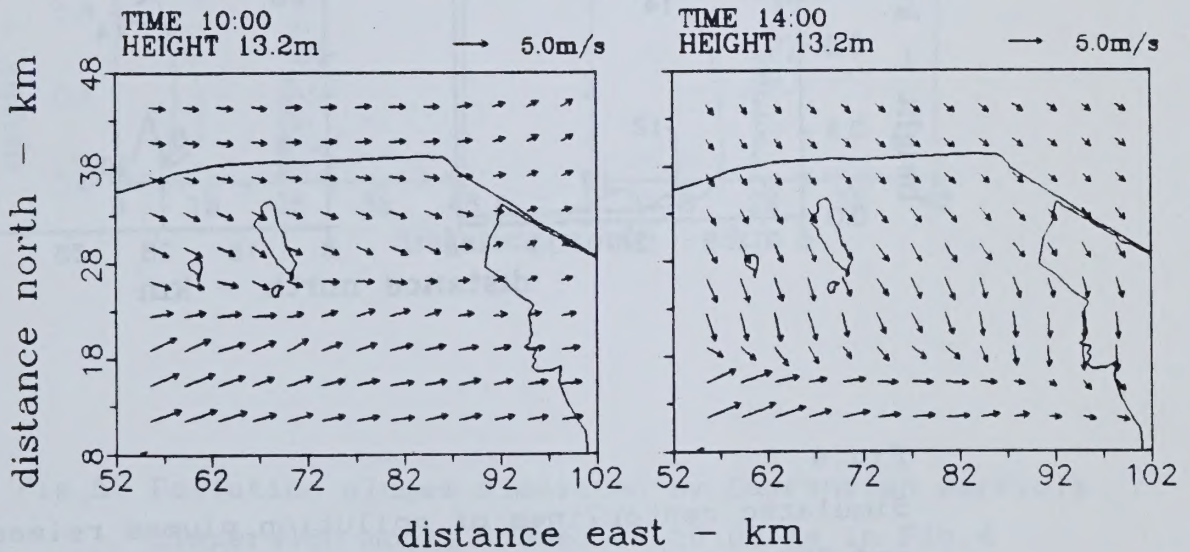


Fig.3. Simulated surface wind fields in the Zarnowiec area

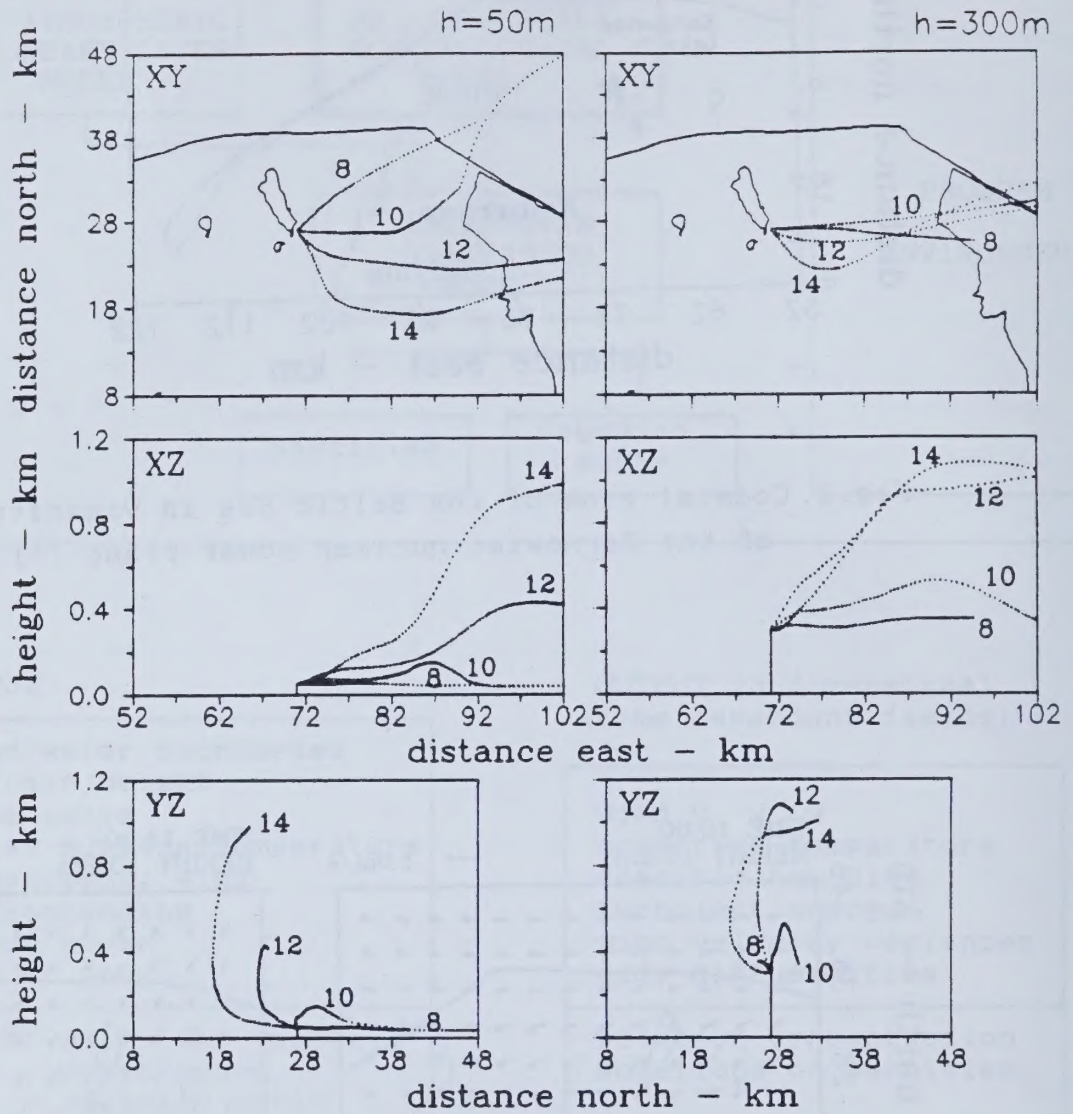


Fig.4.

Simulated centerlines of pollution plumes released from 50m and 300m source at Zarnowiec as viewed in XY section (top), XZ section from South (middle) and YZ section from East (bottom), numbers denote local time

TIME 10:00

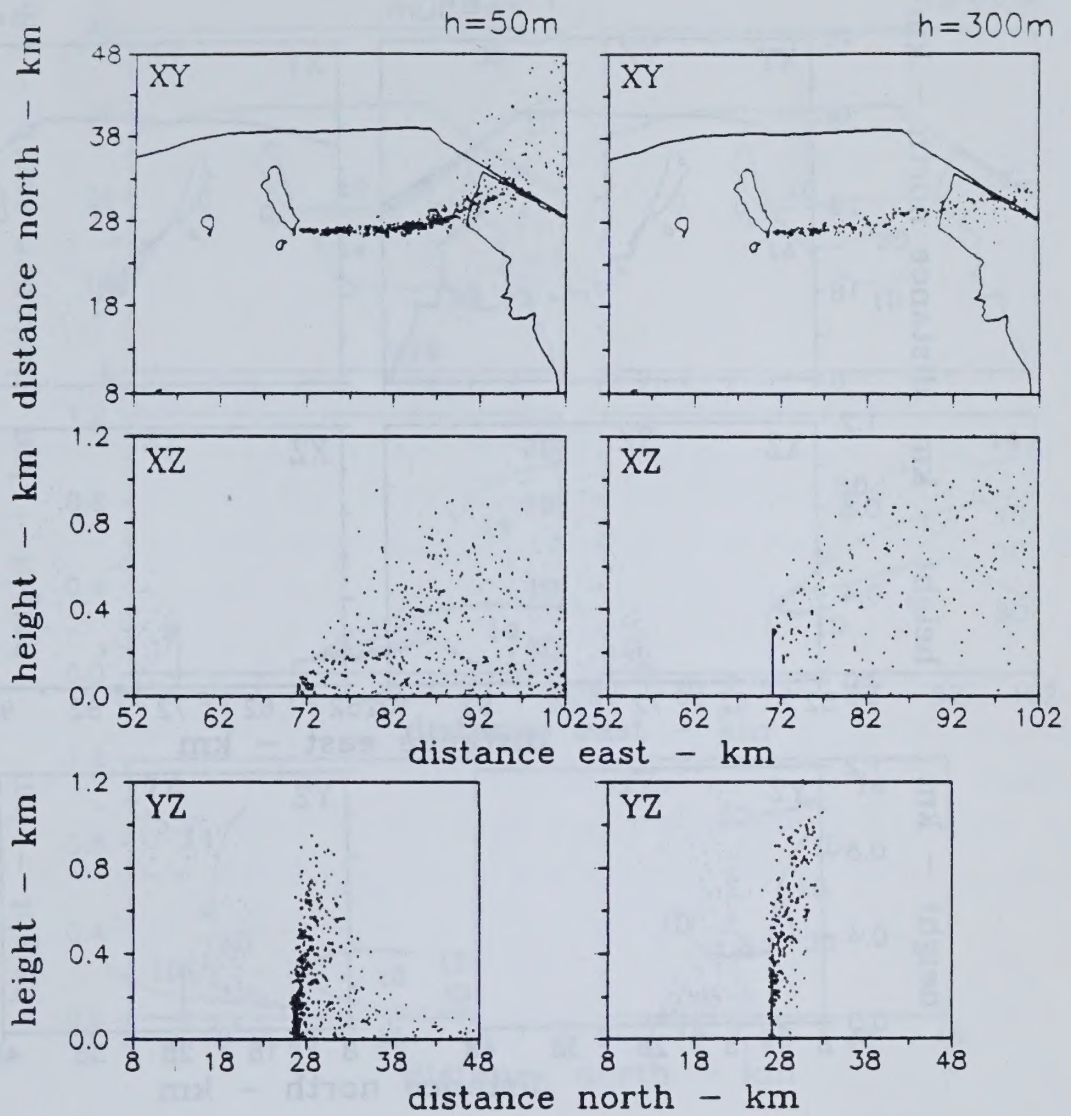


Fig.5. Pollution plumes simulated by Lagrangian particle dispersion model viewed at 10:00 as in Fig.4

TIME 14:00

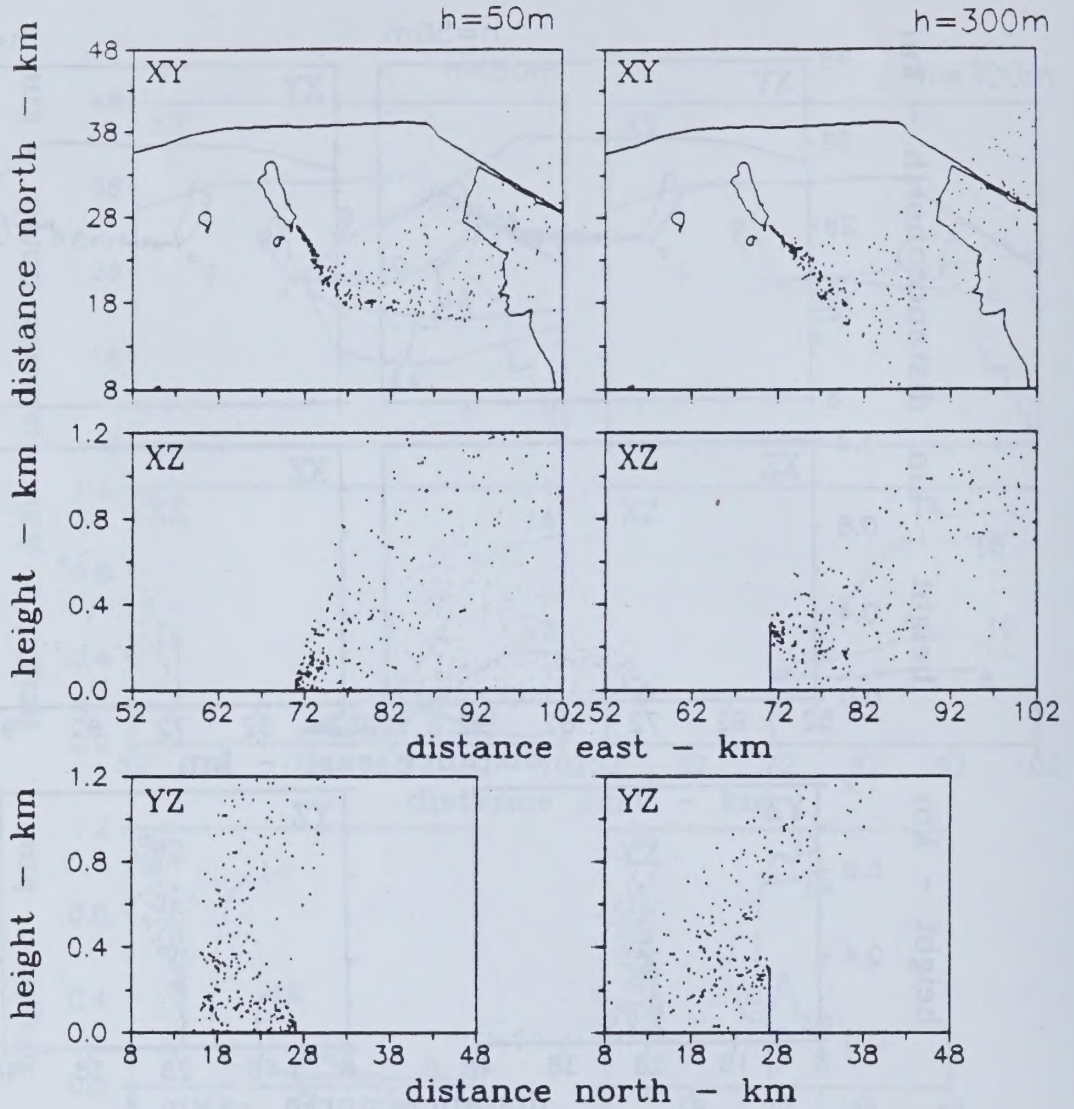


Fig.6. Pollution plumes simulated by Lagrangian particle dispersion model viewed at 14:00 as in Fig.4

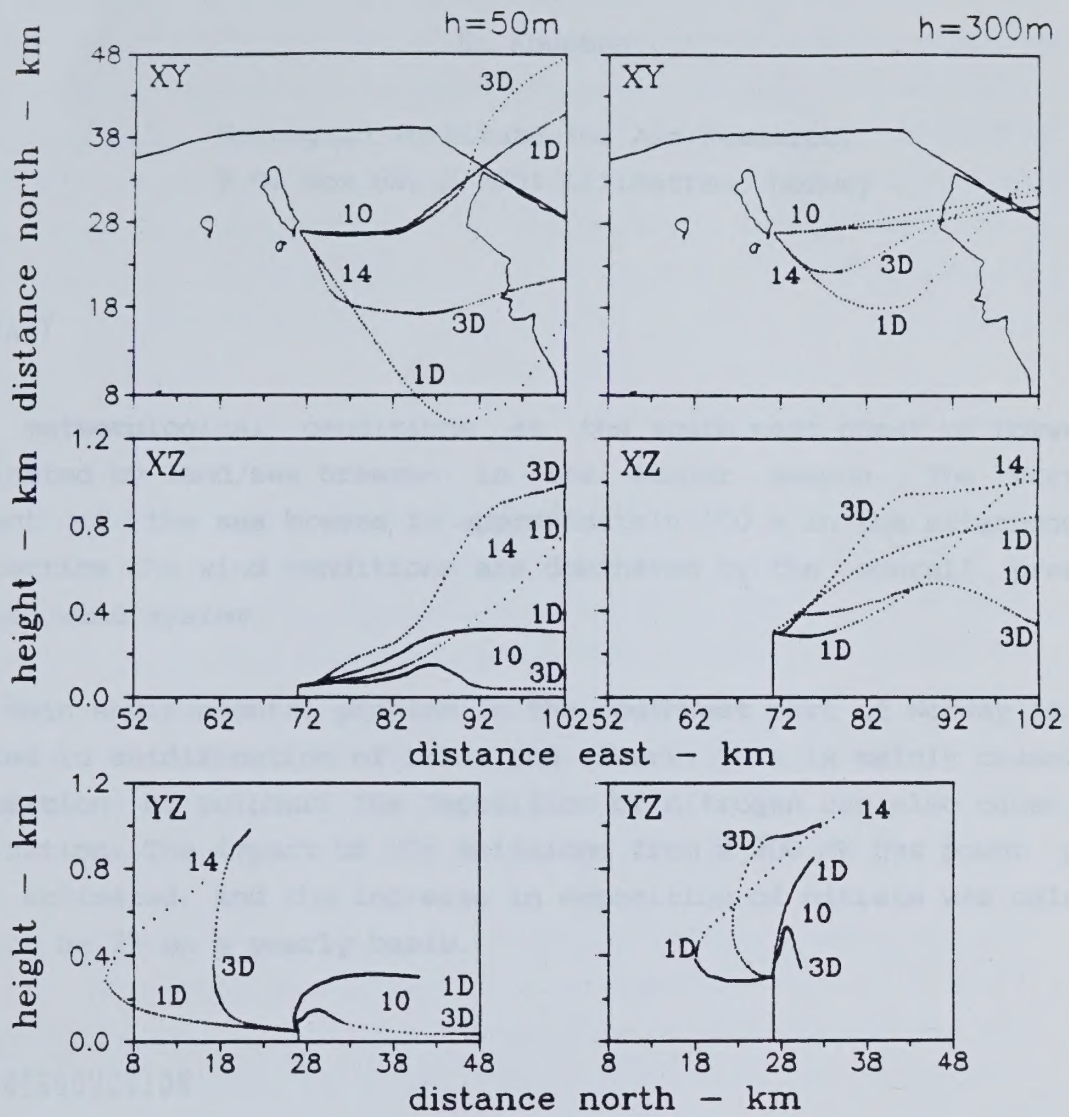


Fig.7. Centerlines of pollution plumes simulated with three-dimensional meteorological fields (3D) and with one-dimensional profiles of meteorological variables taken at Zarnowiec (1D)



Fig. 7. Concentration of pollutant plumes simulated with
 various wind directions. The concentration of pollutant plumes
 is shown in mg/l. The distance north is shown in km. The
 concentration of pollutant plumes is shown in mg/l. The
 distance north is shown in km. The concentration of pollutant
 plumes is shown in mg/l. The distance north is shown in km.

DISPERSION OF NITROGEN OXIDES FROM A GAS POWER PLANT ON THE WEST COAST OF NORWAY

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SUMMARY

The meteorological conditions at the south west coast of Norway is dominated by land/sea breezes in the summer season. The vertical extent of the sea breeze is approximately 500 m in the afternoon. In wintertime the wind conditions are dominated by the overall pressure driven wind system.

The main environmental problem in the southwest part of Norway is connected to acidification of lakes and rivers. This is mainly caused by deposition of sulphur. The deposition of nitrogen can also cause acidification. The impact of NO_x emissions from a 200 MW gas power plant was estimated, and the increase in deposition of nitrate was calculated to be 2% on a yearly basis.

1 INTRODUCTION

At present the electricity demand in Norway is met by hydro electric power. Recently gas fired power plants have been considered, mainly based upon natural gas landed on the western coast of Norway.

The Norwegian Institute for Air Research (NILU) has been involved in estimating dispersion of the air pollutants from these power plants. The sites are located on or near the coast line, where land and sea breezes dominate the wind regimes during the summer season.

The work included collection of appropriate meteorological data by surface mast measurements, SODAR, radiosondes and aircraft measurements.

2 METEOROLOGY

The wind pattern in this part of the country is dominated by a general cyclonic onland wind during summer and an anticyclonic offland wind during the winter season. Overlayed on these large scale winds are mesoscale and local scale onshore and offshore shallow wind systems driven by heating and cooling of the land surface. The mesoscale to local scale wind systems have been investigated on two occasions on 5 June 1987 by radiosondes and on 12 August 1987 by aircraft.

2.1 RADIOSOUNDINGS

Figure 2 illustrate the development of a typical land/sea breeze. On 5 June 1987 four radiosondes were released from Haugsneset (see Figure 1). These releases were at 0600, 0900, 1200, and 1500 GMT. The isotherms in Figure 2 show a rapid surface heating at this specific location between 0900 and 1200. In this period the surface boundary layer reached a height of 500 m. The wind speed gained in strenght during the same period. During the next 3 hours the sea breeze maintained its strenght and has reached a quasistationary state.

The wind shear at the top of the sea breeze was observed at approximately 500 m height. There is a change in wind direction of 90° , from the north-northwest to north-northeast. The atmosphere above the boundary layer was in neutral conditions up to approximately 1800 m. Above the neutral layer a stable stratification was observed.

Information about the mixing height is of great importance when modelling the power plant plumes under these conditions. On a routine basis the mixing height was determined by the potential temperature gradient taken from the radio soundings twice a day at Sola, about 40 km south of Haugsneset. From simultaneous observations of temperature profiles at Sola and Haugsneset, the data from Sola were found to be representative for the gas power plant sites.

2.2 AIRCRAFT MEASUREMENTS

The flight pattern for the aircraft measurements is shown in Figure 1. Measurements were performed along traverses at four different heights; 150 m, 300 m, 600 m, and 1200 m above sea level. Four vertical profiles were also taken; at the locations A, KF, B, and E. Figure 3 shows the isotherms at 0900 and 1700 GMT on 12 August 1987. The thermal boundary layer is seen to build up over land with a smaller build up over the island of Karmøy.

2.3 SODAR-MEASUREMENTS

A SODAR was located at Karmøy (see Figure 1). These measurements show that on an average land/sea breeze is dominating in the summer time. Winds from north-northwest occurred in 25% of the time.

A maximum in turbulence intensity was found during winds from north-northeast. In this sector the largest roughness elements are located. The turbulence intensity decreased with height up to approximately 100 m and was fairly homogeneous with wind direction from this height up to 200 m. The maximum turbulence for winds from north-northeast can be seen even at 200 m, which was the highest height considered.

3 OZONE CONCENTRATION AND CHEMICAL MODELLING

A gas power plant emits NO_x , 85-95% of the nitrogen oxides appears as nitric oxide (NO) and the rest as nitrogen dioxide. For the conversion of NO to NO_2 and further to NO_3^- (approximately up to 100 km from the source), the concentration of ozone in the ambient air is critical. The ozone in the ambient air were measured at ground level and by aircraft. The NO_x -concentrations were monitored at ground level for two years.

The measurements show that the ozone concentrations were typically $60\text{--}70 \mu\text{g}/\text{m}^3$ in daytime during summer. There were no particularly high level episodes during the measurement period. The highest concentration measured was $135 \mu\text{g}/\text{m}^3$.

To estimate the conversion of NO to NO_2 and further to NO_3^- a simple chemical model was built into a puff-trajectory model. The chemical model consists of 4 reactions and the mixing of the individual puff were assumed to be instant. The chemical reactions can be seen in Table 1.

Table 1: Chemical reactions.

R1:	$\text{NO}_2 + h\nu \rightarrow \text{NO} + \text{O}_3$	$0-6.2 \cdot 10^{-3}$	s^{-1*}
R2:	$\text{O}_3 + \text{NO} \rightarrow \text{NO}_2 + \text{O}_2$	$2.1 \cdot 10^{12} \cdot 1^{-1450/T}$	$\text{cm}^3 \text{ molec}^{-1}$
R3:	$2\text{NO} + \text{O}_2 \rightarrow 2\text{NO}_2$	$1.5 \cdot 10^{-40} \cdot 1^{1780/T}$	$\text{cm}^6 \text{ molec}^{-3}$
R4:	$\text{O}_3 + \text{NO}_2 \rightarrow \text{NO}_3^- + \text{O}_2$	$1.2 \cdot 10^{13} \cdot 1^{-2450/T}$	$\text{cm}^6 \text{ molec}^{-3**}$

* only active in day time (dependent on light).

** only active at night.

4 NUMERICAL MODELLING AND RESULTS

The gas power plant plume have been modelled using a puff-trajectory model including chemical conversion of $\text{NO} \rightarrow \text{NO}_2 \rightarrow \text{NO}_3^-$. The model used hourly input of meteorology and ozone concentrations. The plume was modelled for one summer month and one winter month. The results shown in figure 4 are for December 1987 from a power plant located at Haugsneset. The source characteristics are shown in Table 2.

Table 2: Source characteristics.

Power production	MW	200
Emission NO_x (as NO_2)	kg/h	160
Release height	m	50
Stack diameter	m^3	7.3
Volume flux	m^3/s	$2.1 \cdot 10^6$
Release temperature	K	373

The calculations show that the maximum one hour average NO_2 concentrations will occur at 8 km and not exceed $20 \mu\text{g}/\text{m}^3$. As a monthly mean

the averaged concentration will be less than $1 \mu\text{g}/\text{m}^3$. The dry deposition will come from nitrogendioxides, but because of the low long time averages the dry deposition will be minimal.

Wet deposition is mainly in the form of NO_3^- . NO_3^- is very hygroscopic so during rain the NO_3^- is removed within one hour with a precipitation intensity of 1-2 mm/h. The measurement of simultaneous wind and precipitation showed that the impact of wet deposition mainly was north of the power plant. The wet deposition will have its maximum 50 km from the source due to the chemical reaction rate which is slow. The monthly contribution of NO_3^- converted to nitrogen in the maximum zone will be $1 \text{ mgN}/\text{m}^2$. This maximum zone will change position every month, so the yearly contribution will be $10 \text{ mg}/\text{m}^2$. This is approximately 2% of the total wet deposition of nitrate today.

There are some lakes in this area where the nitrogen content in water have increased over the last years. This contributes to the acidification of the lakes. The most affected areas are to the south and east of the planned power plant. The ecology in these lakes are strongly affected by acid rain. In areas where the denitrification is not sufficient to absorb all the supplied nitrogen the nitrogen will cause acidification. The contribution of oxidized nitrogen from the gas power plant may have a marginal effect under these circumstances.

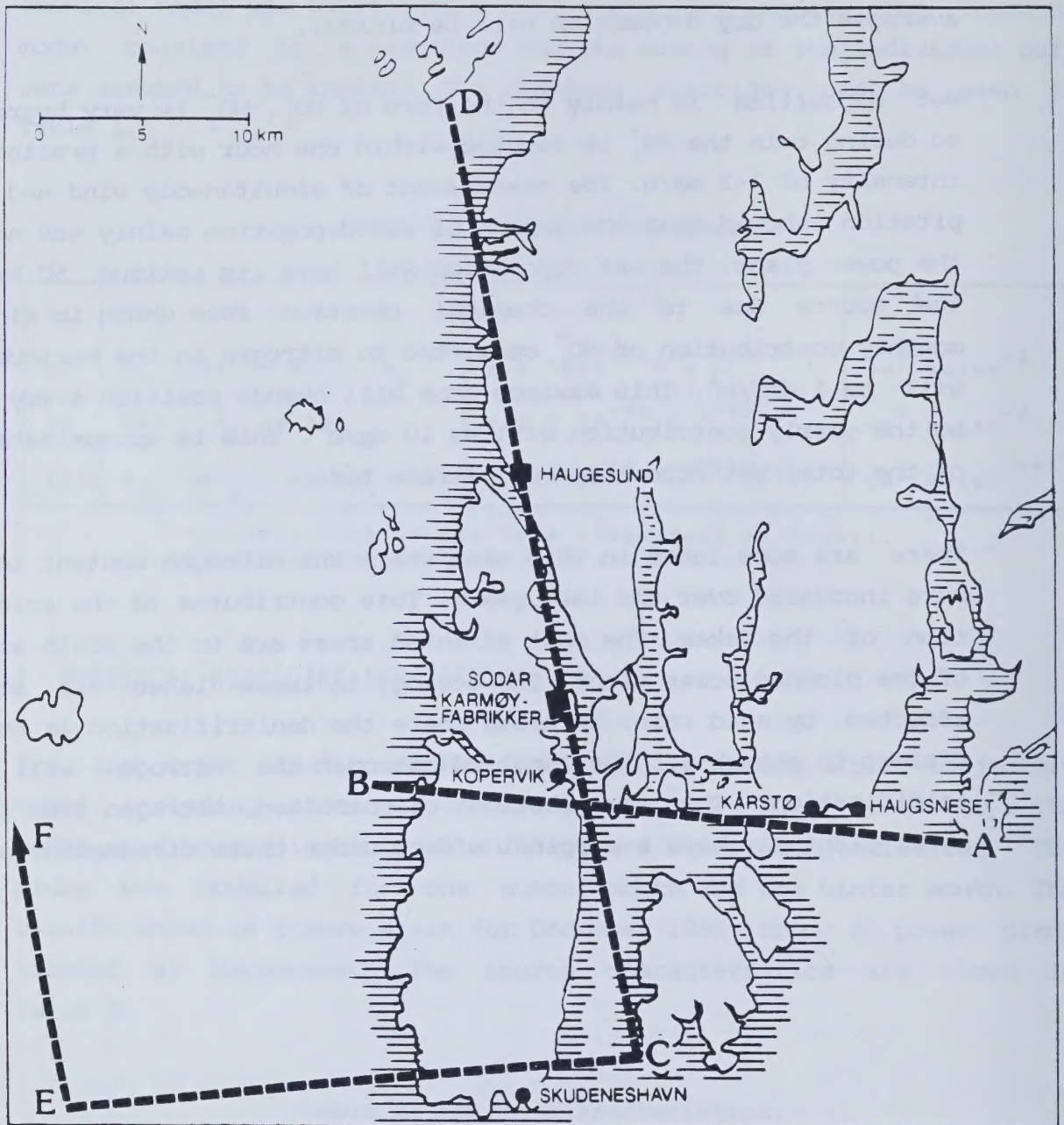


Fig. 1: The location of the power plants with flight plan on the 12 August 1987.

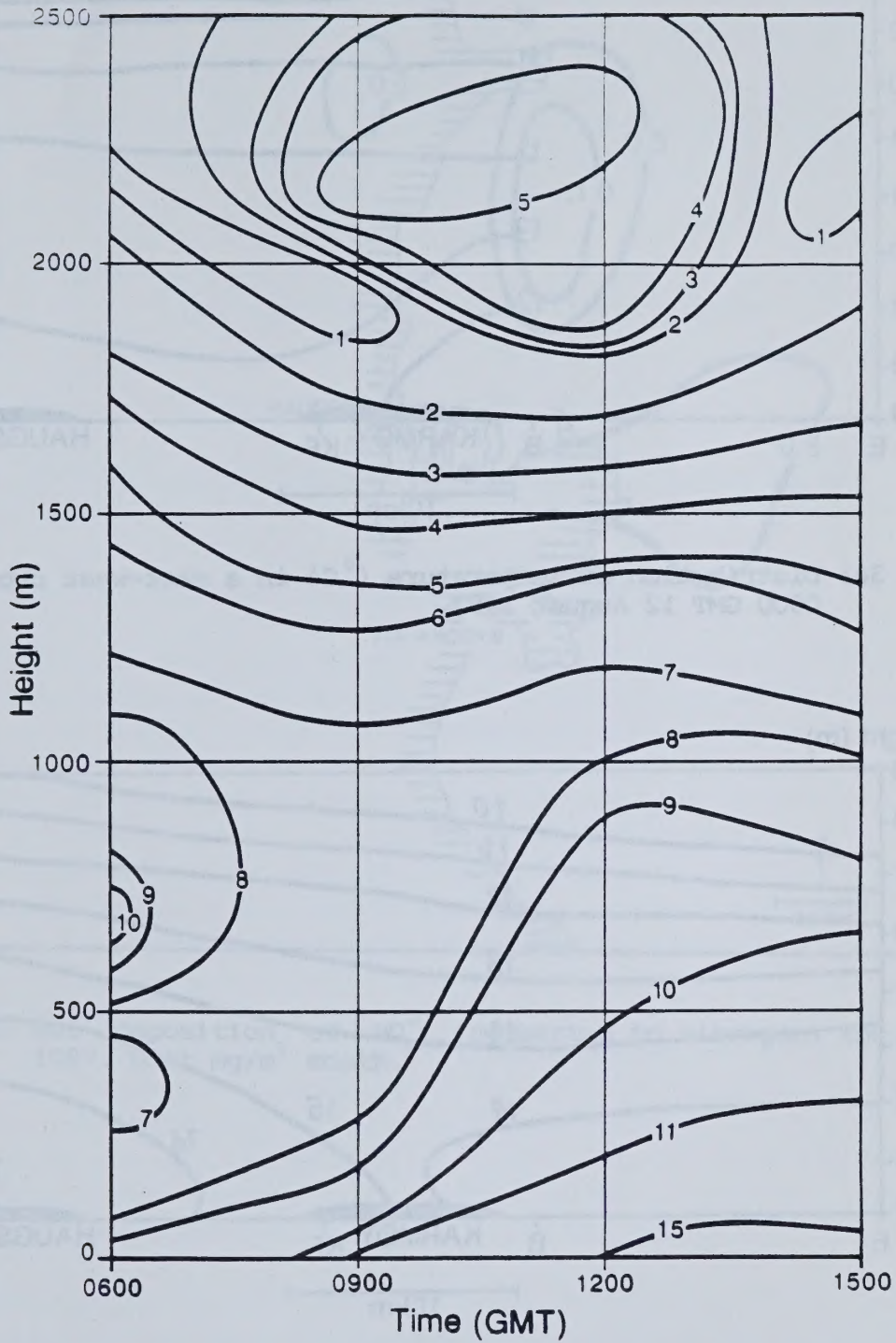


Fig. 2: Vertical temperature ($^{\circ}\text{C}$) distribution versus time at Haugsneset 5 June 1987.

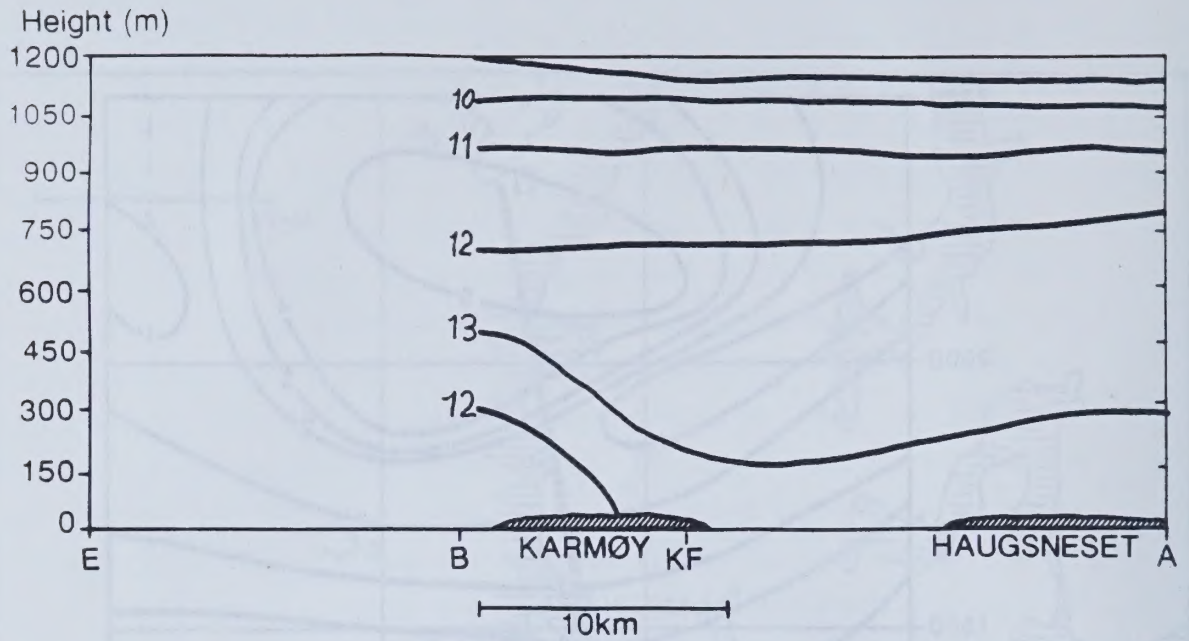


Fig. 3a: Distribution of temperature ($^{\circ}\text{C}$) in a east-west cross section 0900 GMT 12 August 1987.

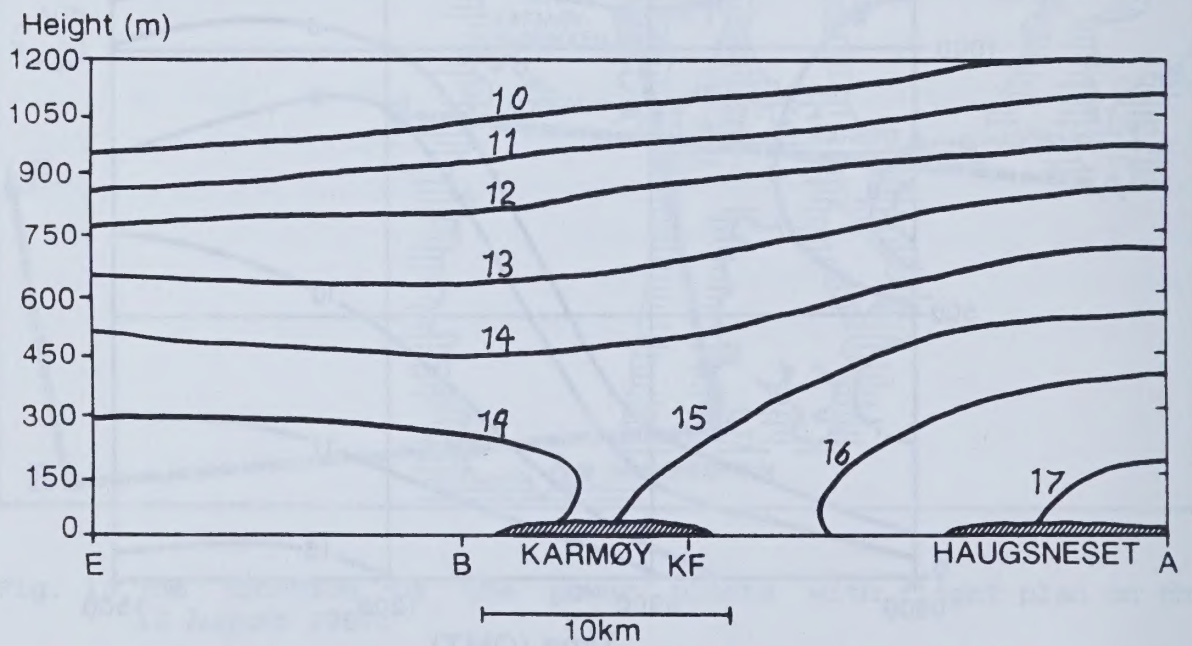


Fig. 3b: Distribution of temperature ($^{\circ}\text{C}$) in a east-west cross section 1600 GMT 12 August 1987.

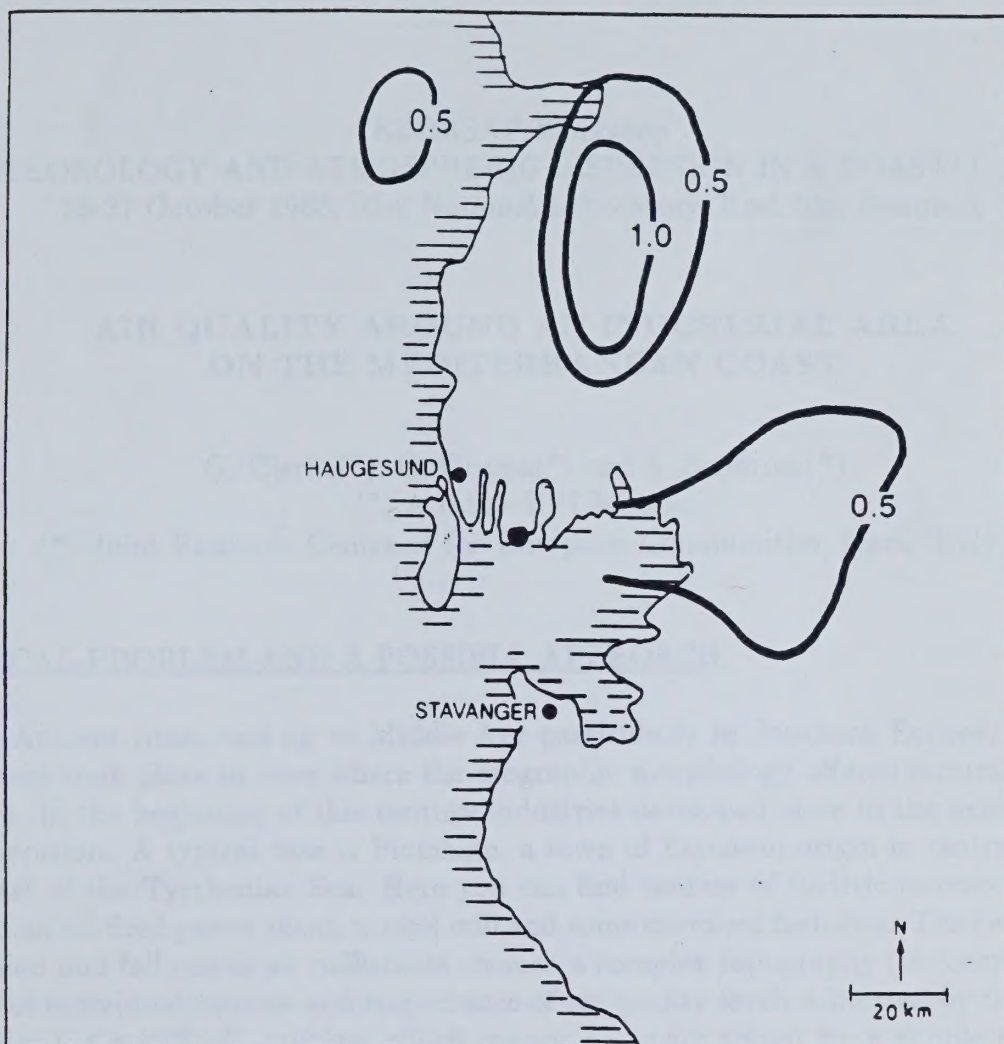


Fig. 4: Wet deposition₂ of NO_3^- (converted to nitrogen) for December 1987. Unit $\text{mg/m}^2 \text{ month}$.

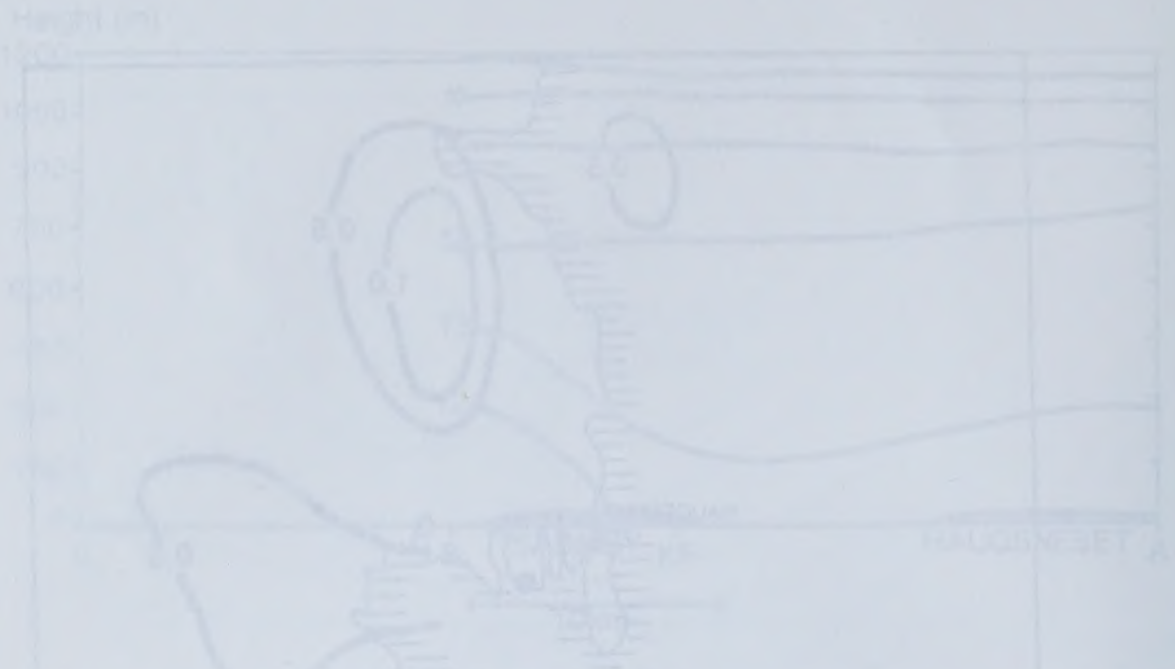


Fig. 2a. Distribution of temperature ($^{\circ}\text{C}$) in a cross-section of the ice sheet (see Fig. 1 for location) on 11 August 1987.



Fig. 2b. Distribution of temperature ($^{\circ}\text{C}$) in a cross-section of the ice sheet (see Fig. 1 for location) on 11 August 1987.

EURASAP Workshop

"METEOROLOGY AND ATMOSPHERIC DISPERSION IN A COASTAL AREA"

25-27 October 1988, Risø National Laboratory, Roskilde, Denmark

**AIR QUALITY AROUND AN INDUSTRIAL AREA
ON THE MEDITERRANEAN COAST**G. Clerici([°]), G. Grippa([°]) and S. Sandroni(^{*})([°]) A.R.S. - ENI Milano;(^{*}) Joint Research Centre of the European Communities, Ispra (Italy)A TYPICAL PROBLEM AND A POSSIBLE APPROACH.

In Ancient times and up to Middle Age particularly in Southern Europe, the urban settlement took place in sites where the orographic morphology offered natural defensive barriers. In the beginning of this century industries developed close to the existing urban agglomeration. A typical case is Piombino, a town of Etruscan origin in central Italy, on the coast of the Tyrrhenian Sea. Here you can find centers of turistic interest which are close to an oil-fired power plant, a steel mill and some chemical factories. The evaluation of dispersion and fall-out of air pollutants on such a complex topography (in terms of contribution of individual sources and respectance of air quality levels admitted by the national legislation) is a difficult problem which cannot be easily solved by a simple monitoring network.

An environmental impact study was required by regional authorities when the electricity board expressed an intention of doubling the capacity of the power plant while changing at the same time the fuel, from oil to coal with a lower S-content. It was therefore necessary to analyse the actual fall-out configuration in terms of contribution of individual sources (industrial and urban) and to simulate the impact of each possible future modification on that complex topography.

For the initialization of this study hourly data of meteorological and chemical parameters extended over more than 10-years have been made available to us. An analysis of the data allowed the identification of typical situations responsible for an appreciable fall-out. Later some short-term intensive campaigns of meteorological sounding were performed in different seasons of the year for a better characterization of the air circulation. Supplementary meteorological informations has been available for the coastal site of Montalto di Castro, about 100 km S from Piombino, where a nuclear plant was in construction (Cagnetti et al.,1984). The emission inventory has also been made available for this study.

For the modelisation of the identified situations leading to pollution accumulation we applied an Eulerian grid model. It was previously developed for the reconstruction of

the wind-field in a complex terrain (Fos-Berre, France; Clerici and Sandroni, 1985) and modified by the introduction of pollutant transport equations in the previous set. By this model we could reconstruct the air masses circulation and compute the fall-out values to be compared with the admitted limit values. For typical atmospheric circulation patterns it was possible to map for each source, real or potential, the corresponding iso-deposition lines above the considered area.

THE AREA AND ITS LOCAL CIRCULATION

The area under consideration is shown in Fig. 1: at W the Tyrrhenian sea and the island of Elba and at E the pre-Appennins (Monte Scodella, 335 m and Monte Calvo, 465 m, up to Scarlino e Massa Marittima). The natural barrier of Piombino towards the sea is Monte massoncello, 286 m. Two small flat areas with a triangular shape can be identified, Piombino-Venturina-Vignale and Follonica-Massa Marittima-Portiglione. Over these two flat areas, where towns and industries are sited, the air circulation is strongly constrained by the orographic barriers and channelled along the main direction San Vincenzo-Torre del Sale, Suvereto-Torre del Sale and Valpiana-Follonica. Also the sea breeze regime and the urban heat islands contribute in modifying the local circulation.

The climatology and meteorology on a synoptic scale is typical for the Western Mediterranean area. From Autumn to mid Spring the highest frequency of surface cyclons per unit area of the entire Northern hemisphere occurs, while the zonal type circulation is rare (Castro, 1988). This fact is a consequence of the thermal contrast between sea and land or of the invasion of cold air masses from higher latitude. Typical surface circulations are NW-, SW- or E-flow, for which tall stack plumes move parallel to the coast and to the mountains or offshore. During the Summer, cyclons are very rare and the land is warmer than the sea. The area enjoys clear skies and high insolation interrupted only by local storms.

The critical conditions for strong deposition inside the area are associated to a high pressure field at the synoptic scale and by weak air circulation at ground level. This fact already observed at Montalto di Castro (Cagnetti et al., 1984 and 1988), Cartagena and Castellon (Millan et al., 1988) increases the importance of local scale phenomena, such as the breeze regime. In particular the sea breeze and the transition sea-to-land breeze and viceversa are the critical situations for increased fall-out. These breeze situations are present also in the cold season, but its contribution is strongly attenuated. In the transition seasons perturbed weather conditions alternate with prevailing instability or high-pressure fields. Generally, the sea-breeze starts around 9.00 GMT and reaches its maximum around 13.00 GMT. Late in the afternoon it is converted to the weaker land-breeze. Due to the presence of orographic barriers and urban heat islands the air circulation is strongly modified, particularly at ground level and the few stations distributed inside the area are insufficient to describe the wind field.

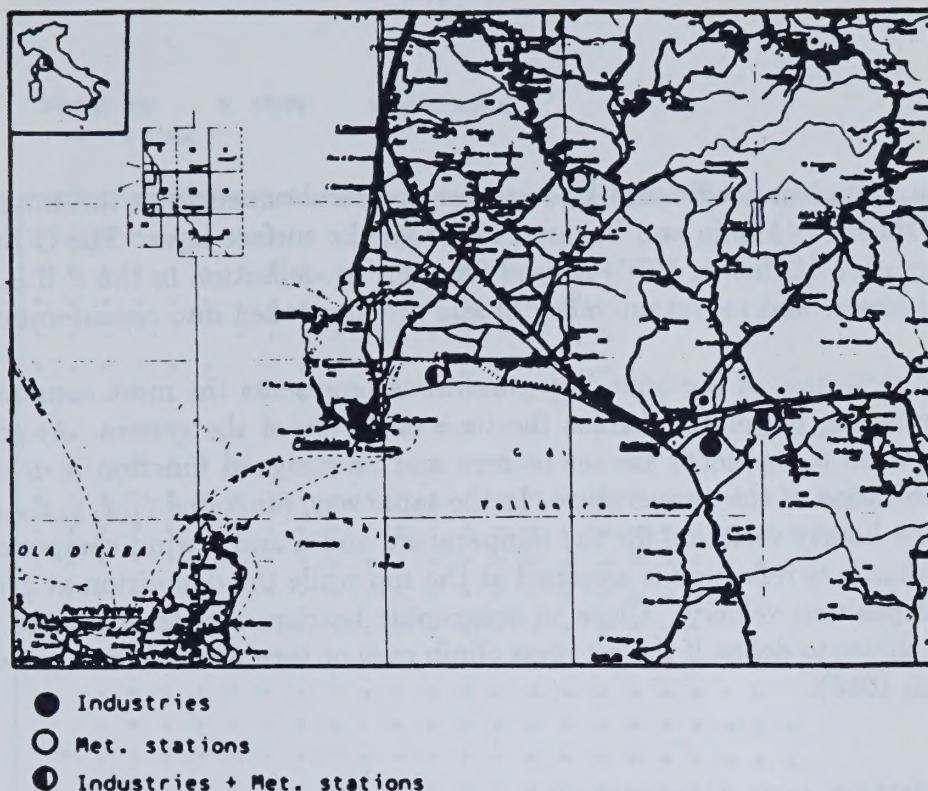


Fig. 1 - Map of the area under consideration (60x40 km.)

THE MODEL

For the reconstruction of the three-dimensional wind field and fall-out distribution we referred to a time-dependent Eulerian grid model previously developed for a complex topography situation (Clerici and Sandroni, 1985). The model is based on the Navier-Stokes equations completed with the equation for pollutant transport. This set of equations derived with the hydrostatic approximation and the Boussinesq turbulent closure, can be written as follows:

$$\frac{\delta u}{\delta t} + u \frac{\delta u}{\delta x} + v \frac{\delta u}{\delta y} + w \frac{\delta u}{\delta z} = f(v - v_g) + \frac{\delta}{\delta z} (k_m \frac{\delta u}{\delta z})$$

$$\frac{\delta v}{\delta t} + u \frac{\delta v}{\delta x} + v \frac{\delta v}{\delta y} + w \frac{\delta v}{\delta z} = f(u_g - u) + \frac{\delta}{\delta z} (k_m \frac{\delta v}{\delta z})$$

$$\frac{\delta u}{\delta x} + \frac{\delta v}{\delta y} + \frac{\delta w}{\delta z} = 0$$

$$\frac{\delta T}{\delta t} + u \frac{\delta T}{\delta x} + v \frac{\delta T}{\delta y} + w \frac{\delta T}{\delta z} = \frac{\delta}{\delta z} [k_h (\frac{\delta T}{\delta z} + \Gamma)] - w\Gamma$$

$$\frac{\delta c_i}{\delta t} + u \frac{\delta c_i}{\delta x} + v \frac{\delta c_i}{\delta y} + w \frac{\delta c_i}{\delta z} = \frac{\delta}{\delta z} (k_p \frac{\delta c_i}{\delta z}) + S_i$$

where u , v and w are the horizontal and vertical wind field components, u_g and v_g are the horizontal geostrophic wind components defined by the Hess approximation (Hess, 1968), T is the temperature, c_i is the concentration of the pollutant i , f is the Coriolis parameter and Γ the dry adiabatic lapse-rate.

The turbulent coefficients k_m , k_h and k_p are obtained from the similarity theory of Monin-Obukhov (Monin and Yaglom, 1971) for the surface layer. The O'Brien polynomial approximation (O'Brien, 1970) is used for their modelisation in the P.B.L. Effects due to sea-land breeze and mountain-valley breeze are also taken into consideration.

The definition of the boundary conditions represents the most constrainig aspect for the calculation, because it defines the time evolution of the system. At ground (z_g), the u and v wind components are set to zero and an assigned function is used to define the hourly variation of the temperature. In the same way, functions $f_j(x, y, z_{top}, t)$, are used to obtain the hourly variation for the temperature and u and v wind components at the top. For pollutants, a reflection is assumed at the top while the deposition at ground is defined via its deposition velocity. Close to orographic barriers a criterion based on the Froude number allows to define if the air mass climb over or turn around the obstacle (Clerici and Sandroni, 1985).

RESULTS AND DISCUSSION

The reconstructed wind fields at a reference altitude (200 m.) are shown in Fig. 2 respectively for sea-breeze, land-breeze and transition situations. In the first two cases the circulations around Monte Massoncello, the channeling effect along Torre del Sale-San Vincenzo direction and stagnant air in the Follonica-Scarlino plain are evidentiatiated while the transition situation shows a weak and variable wind field in the two plains Piombino-Venturina-Vignale and Follonica-Scarlino.

In Fig. 3 the computed isolines for SO_2 ground concentration (referred to the same situation of Fig. 2) are given. The highest ground concentration appears in sea-breeze and transition conditions in the Scarlino-Follonica area, where stagnant air is observed. On the contrary the town of Piombino has relatively high ground concentration on night, in land-breeze conditions.

In Fig. 4 the daily evolution of SO_2 and particles ground concentration for two different locations are shown. The evaluation of this parameter for particulates is very difficult for the experimental point of view. At Scarlino (Fig. 4a), highest values are observed on sea-breeze condition, the different peaks corresponding to different sources. At Piombino town (fig. 4b), the SO_2 maximum appears on the transition sea-to-land breeze and it is associated to the Torre del Sale power plant. From 22.00 to 6.00 GMT during land breeze particulates due to still mills reache the highest values ($54 \mu\text{g}/\text{m}^3$); lower peaks at around 20.00 GMT and 10.00 GMT are associated to breeze transition and are mainly related to one of the steel mill.

In conclusion, the applied model succedeed in describing the air circulation and its time evolution on complex terrein. It has also allowed to describe the breeze transitions, which are considered by a general feeling, as critical situations for increased fall-out.

Summer day h: 13:00 height: 200 m
($\rightarrow$ = 5 m/sec)

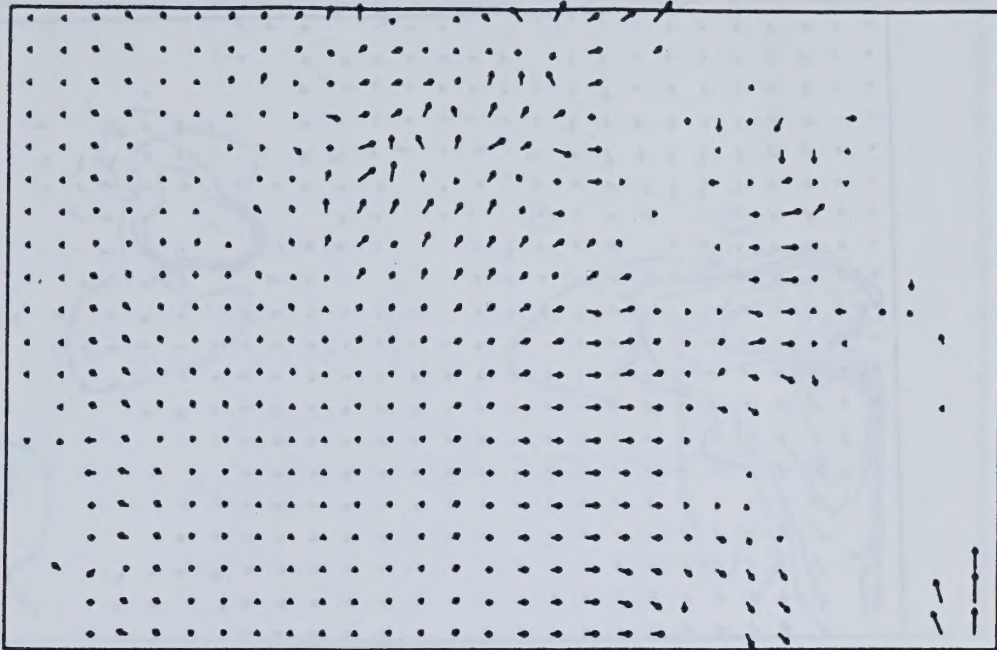


Fig. 2a: wind field in sea-breeze conditions

Summer day h: 21:00 height: 200 m
($\rightarrow$ = 5 m/sec)

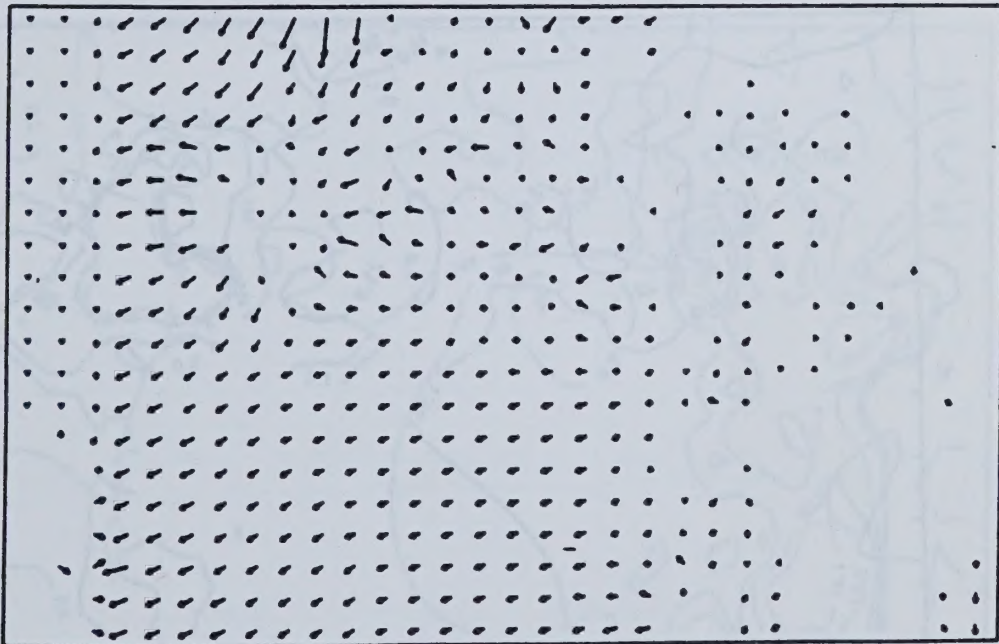


Fig. 2b: wind field in land-breeze conditions

Summer day h: 09:00 height: 200 m
 ($\rightarrow$ = 5 m/sec)

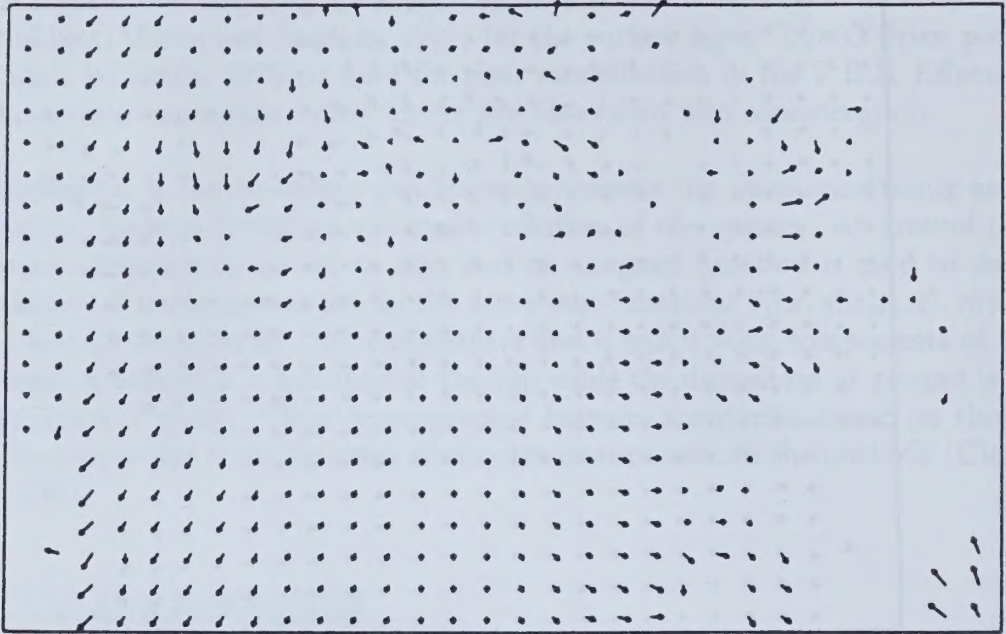


Fig. 2c: wind field in transition conditions

Summer day h : 13:00
 Concentration of SO₂ at the ground ($\mu\text{g}/\text{m}^3$)

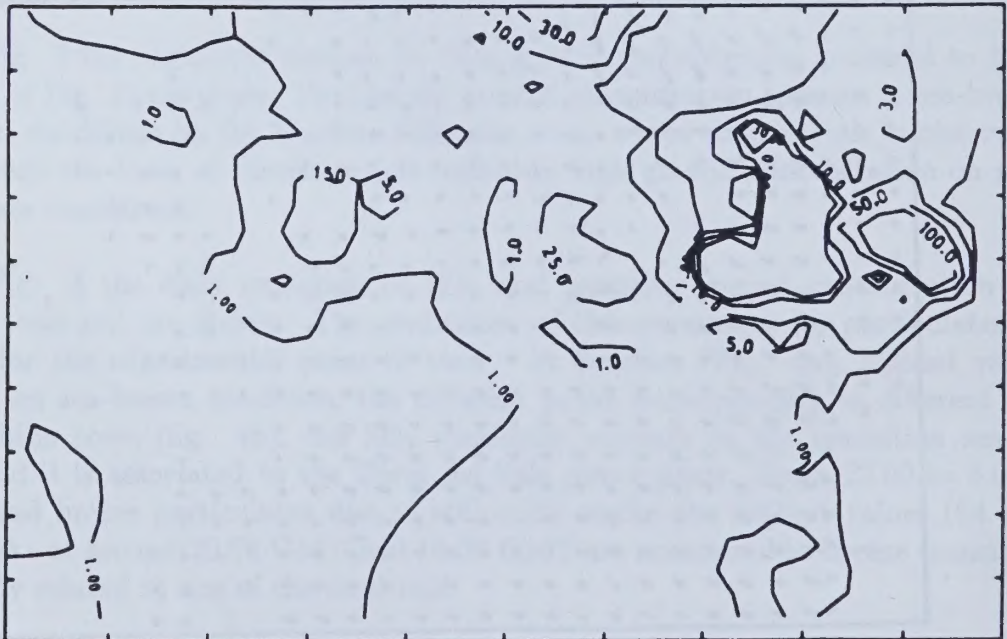


Fig. 3a: isolines relative to sea-breeze conditions

Summer day h : 21:00
 Concentration of SO₂ at the ground (µg/m³)

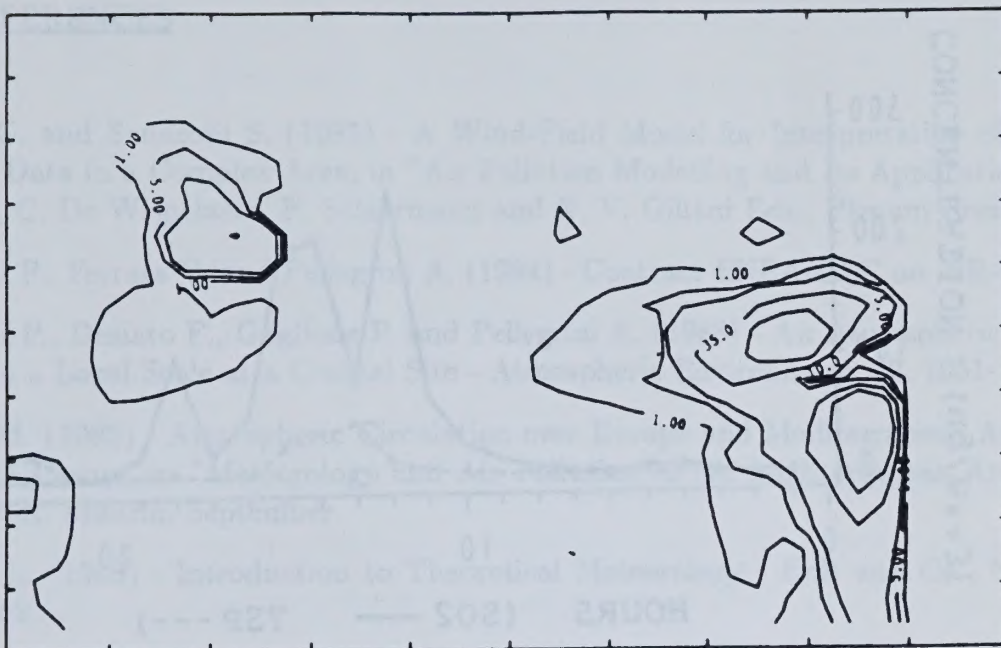


Fig. 3b: isolines relative to land-breeze conditions

Summer day h : 09:00
 Concentration of SO₂ at the ground (µg/m³)

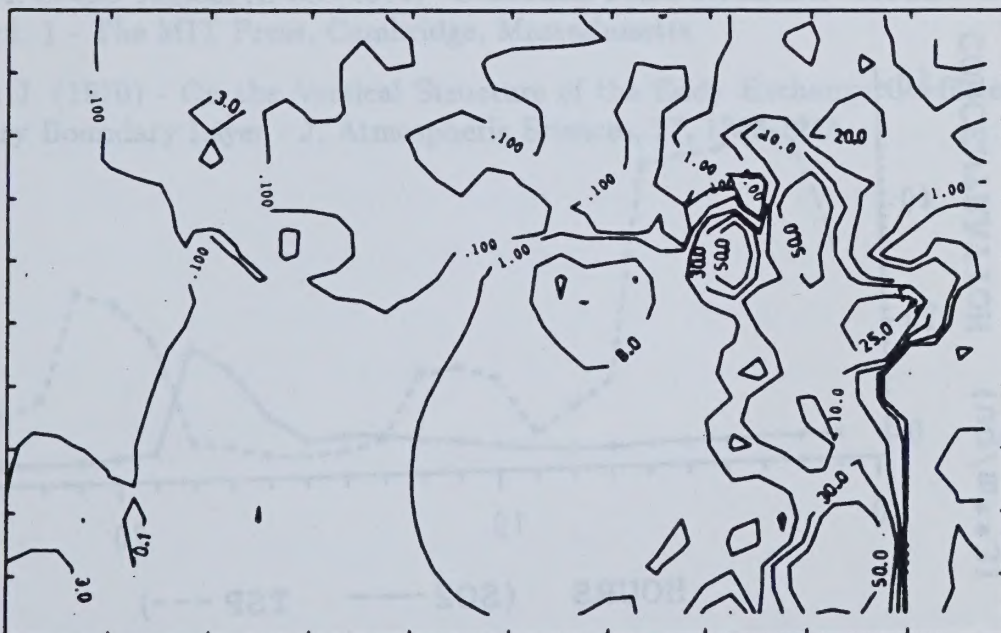


Fig. 3c: isolines relative to transition conditions

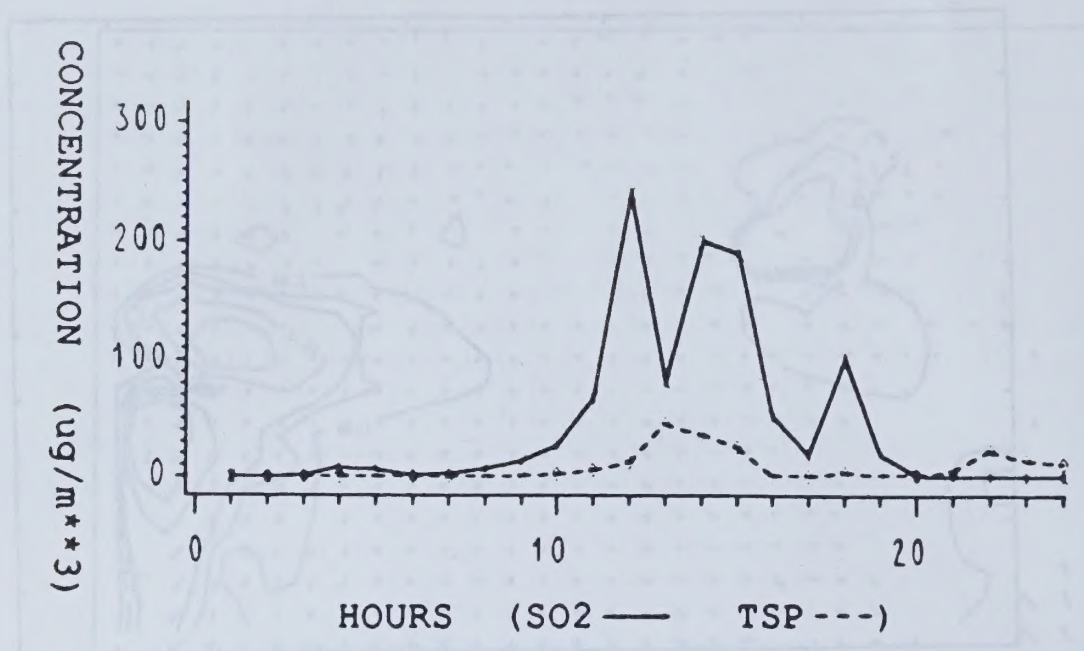


Fig. 4a: hourly ground concentration at Scarlino

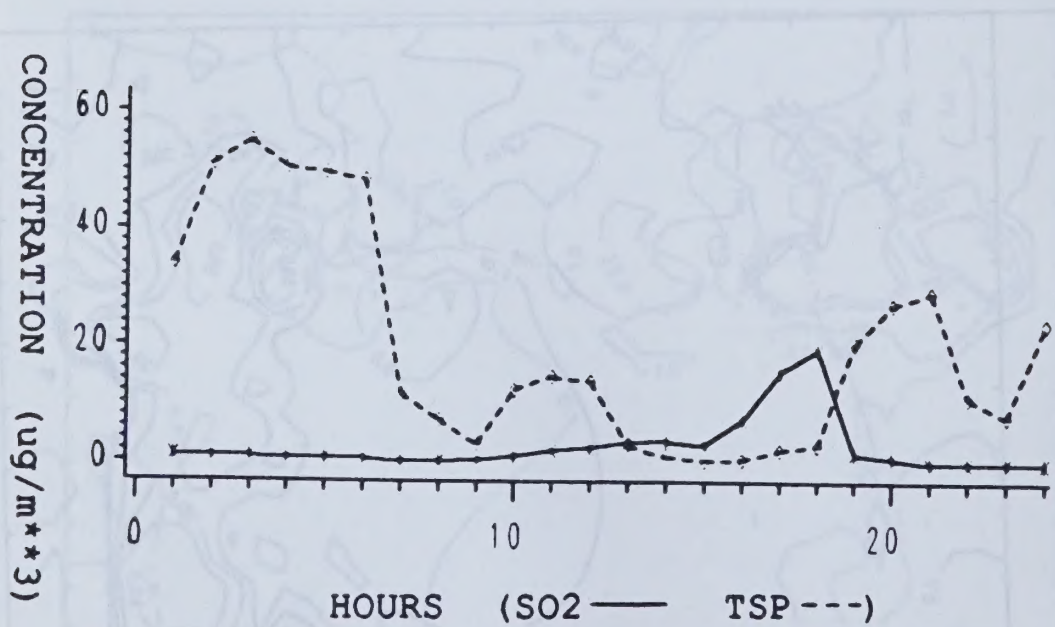


Fig. 4b: hourly ground concentration at Piombino

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PERFLUOROCARBON TRACER EXPERIMENTS IN
A LAKE-MOUNTAIN AREA (CAMPO DEI FIORI EXPERIMENT)

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1. Introduction

This paper describes some results from a full scale tracer experiment in which the atmospheric dispersion in a lake-mountain area was investigated. The experiments were carried out under the condition of a cold lake surface, and a well developed anabatic wind on the heated slope of the mountain. The height difference between the lake surface and mountain ridge is about 800 m. The tracer was released upwind of the lake under very unstable atmospheric conditions, with light wind. The measurements show that the effect of the cold lake surface delays the arrival of the tracer plume to the downwind lake-shore; a strong channeling effect on the plume is created by the mountain and the tracer concentration on the mountain ridge and leeward is affected by a strong vertical wind component before the ridge. An attempt to simulate the dispersion by a segmented plume model failed completely.

In these experiments, use was made of the perfluorocarbon tracer technique. Considering the novelty of this tracer technique it is described in some details.

2. Tracer and meteorological measurements

2.1 Release. The perfluorocarbon tracer (perfluoromethylcyclohexane = $C_{7F_{14}}$) is available in the liquid phase. The release is carried out by atomization and complete evaporation in a gas stream.

The liquid is withdrawn from a storage tank and then forced by a pressure metering pump through a nozzle generating a liquid aerosol. This spray is then injected into a warm air stream produced by a blower followed by a heater. The release rate of the tracer is determined by the diameter of the nozzle and the pressure of the liquid.

The air stream and its temperature must be adjusted to assure rapid and complete evaporation of the tracer, according to the chosen release rate. The total quantity of tracer released is determined by weighing before and after the discharge.

2.2 Sampling. Automatic sampling units are used to collect the samples of air that are going to be analysed for tracer content. The air is sampled into plastic bags made of SARAN (trade name for polyvinylidene chloride). The automatic sampling units allow sequential sampling in up to eight bags. They can be programmed by an external computer to a preselected waiting time before starting and individual sampling times are used for each bag. The sampling units are equipped with a membrane pump and an electronic system which secures constant airflow to the bags. The sampling units are battery operated (30 hours) and waterproof for field use.

2.3 Analysis. The determination of the perfluorocarbon tracer concentration in air were performed by means of the Dual-Trap Perfluorocarbon Analyzer originally conceived and developed by Lovelock and later modified by the Brookhaven Laboratory. It can detect perfluorocarbon tracers in air samples at concentrations around 10^{-2} ppt (Ref. 1). The Dual-Trap analyzer has two ambersorb traps: while one is sampling, the other is thermally desorbed and the air led to a chromatograph column that is followed by an electron capture detector that determines the tracer content. The time required for one measurement of the perfluorocarbon tracer (C_7F_{14}) at 10^{-2} ppt level is about 5 minutes.

2.4 Comparing SF_6 and C_7F_{14} . The reliability of the tracer technique described above has been investigated in the laboratory and during field conditions (Ref. 2). Here we report only the results obtained during the experiment performed on July 12 in which the two tracers, SF_6 and C_7F_{14} , were released and sampled simultaneously.

Figure 1 shows the normalized ratio of the tracer concentrations. In this experiment the ratio of the release rate of the two tracers was 22.9, while the mean ratio between the measured SF_6 and C_7F_{14} concentrations, based on 21 samples, resulted in 23.5 ± 5.0 .

2.5 Meteorological measurements. Within the experimental area wind speed and direction near ground level was measured at 3 locations: at the release point to obtain the boundary conditions, 2 km NNW of the lake in the valley west of the Campo dei Fiori mountain to measure the direction and strength of the channeled air, and at the meteorological observatory at the ridge of Campo dei Fiori to measure the conditions on the mountain.

Furthermore vertical profiles of temperature and tracer concentration was measured by a helicopter at selected locations (near tracer release point, over the lake, downwind lake-shoreline, slope on the mountain, ridge of the mountain and downwind the mountain, see Fig. 2).

At the Ispra-establishment 10 km west of the area, sodar and sonic-anemometer measurements were carried out, giving information on the wind profile and turbulence characteristics.

3. Experimental plan

Three experiments were carried out on: July 12, 19 and 21, 1988. During the first experiment two tracers (SF_6 and C_7F_{14}) were released and sampled simultaneously in order to compare the results, in the remaining experiments only the tracer C_7F_{14} was used. The release height was about 2 meters. The 45 sampling units were positioned in five arcs between 1.5 km and 12 km downwind from the release point. Three additional sampling points were located on the top of another hill at a

downwind distance of about 15 km. The automatic sampling units collected 6 samples, each with 30 min. sampling period. Measurements of vertical tracer and temperature profiles were carried out at seven locations by a helicopter. Figure 4 shows the map of the sampling network.

The table summarizes the scheduled plan in Central European Time for the tracer release and sampling.

4. Discussion of experimental results

In the following we concentrate the discussion on the experiment of July 21. This experiment was chosen, because the tracer sampling unit set-up covers the tracer plume better than in the two other experiments, and the mountain wind was relatively strong. All conclusions have been verified against the experiments on July 12 and 19.

4.1 Temperature field. Gross features of the temperature field above the area were constructed from the temperature measurements carried out by the helicopter. Figure 2 illustrates the 20, 23, 25 and 27°C isolines of temperature (isotherms). The effect of the cold lake-surface is seen by the lowering in height above the lake of the 27°C isotherm, whereas the height of the other isotherms remains largely unchanged. The heating of the southern mountain slope by the sun causes the height of the isotherms to increase giving rise to the anabatic wind that blows along the slope. On the northern, unheated, slope and further downwind, the isotherms return to the height before the influence of the lake and mountain was felt in the temperature field. It is interesting to note that the maximum height of the isotherms are actually reached before the ridge of the mountain. Between the maximum point and the ridge the height of the isotherms decreases rapidly, and then remains nearly constant further downwind.

Figure 3 illustrates details of the temperature conditions over the lake. Upwind, the air is unstable with a sharp decrease in potential temperature between the ground and 50 m. Over the cold lake the air is slightly stable/neutral stratified and approximately 1°C

colder than the air over land. The height of the stable layer is about 200-300 m near the northern shoreline. The potential temperature profile measured only 1 km inland downwind of the lake shows that the stable air over the lake is already convective and heated by several degrees.

4.2 Dispersion. Studying Fig. 4 several effects on the dispersion of the tracer plume caused by the lake and mountain can be observed. High tracer concentrations are found already in the first sampling period at the north-western part of the lake. This area is flat, low-level land bordered to the north-east of the Campo dei Fiori mountain. The high concentrations must be caused by channeling of the flow along the mountain.

In the tracer sampling arc at the northern lake shore, the concentrations during the 2 first measuring periods (total 60 min.) are relatively small, and followed by a distinct jump in the third measuring period. This indicates the arrival of the plume. Taking the travel time as 1 h the travelling speed of the plume over the lake is 1 m/s, which is similar to the wind speed (1.1 m/s) measured at the release point. In a similar way the travel time for the plume to the ridge of the mountain can be determined to 1 1/2 h, resulting in a travel speed of the plume of 1.7 m/s.

Comparing the tracer concentrations at the northern lake-shore with the concentrations on the slope of the mountain 300 m higher, it can be seen that during the first 60 min measuring period, the concentrations on the slope 7 km from the release point are higher than near the lake only 4 km from the release-point despite the fact that the tracer was released near the ground. This suggests that the tracer at the slope and at the shoreline follows different routes. A part of the tracer will be embedded into the stable air above the lake surface. The vertical transport of momentum in this layer is inhibited compared to over land conditions, which prevents the wind speed in the stable layer over the lake from increasing in response to the mountain wind, resulting in a low travelling speed of the tracer plume. The phenomena

was already discussed by Doran and Gryning (Ref. 3). The tracer that is measured at the northern lake-shore must pass through the stable layer, either from above - being difficult - or be caught by the layer over the lake and transported within it and therefore the fully developed plume will arrive relatively late. However, another part of the tracer plume will be caught by the mountain wind and pass over the stable layer, over the tracer sampling network at the shoreline and reach the mountain slope causing the relatively high concentrations at the slope. The flow is depicted in Fig. 5.

The tracer concentration measurements carried out by the helicopter (Fig. 6) show that over the lake most of the tracer is expected to be found into the stable layer. When the tracer reaches the northern lake-shore, it is vertically mixed already a few kilometers inland as well as at the mountain slope. At the top of the mountain, the high concentration 500 m above the ridge should be noted. Before reaching the ridge the tracer plume splits into 2 parts, one part passes over the ridge and into the neighbouring valley, the other part continues upward, Fig. 5. The lower part of the concentration profile belongs to the air that passes over the ridge, the upper part to the ascending air-mass.

The above discussion shows that the flow and dispersion processes in this type of terrain is controlled by temperature differences and orography. Therefore, simulations of the wind-field and atmospheric dispersion under these conditions are likely to be successful only when models that include these effects, in physically realistic ways, are applied.

5. Numerical simulation.

The tracer experiment on July 21 has been simulated by a segmented Gaussian dispersion model (Ref. 4). In the model the plume is divided into a series of segments (puffs), where each segment evolves according the local meteorological conditions at the meshpoint it occupies. The

wind-field on the mesh is obtained from the meteorological measurements by a zero-divergence (mass consistent) interpolation technique.

The dispersion of each element along the plume centerline is described by means of vertical and lateral dispersion coefficients which depend on the assumed turbulence of the mesh considered. In the simulation Pasquill category F of stability over the lake was taken surrounded by a strongly unstable region (category A).

From the results of the simulation it is evident that this type of model is inadequate for simulations of the atmospheric wind-field and dispersion over the area.

The channeling of the plume on the western side of Campo dei Fiori is not reproduced; the effect on the travel speed of the plume caused by the cold lake surface is not found, the simulated tracer concentrations do not agree with the measured ones, Fig. 7, especially those concerning the tracer concentrations leeward of the ridge, as the model does not take into account the strong vertical wind component before the crest.

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"CAMPO DEI FIORI" CAMPAIGN - SCHEDULED PLAN FOR THE EXPERIMENTS

RELEASE		SAMPLING				GROUND LEVEL POINTS				HELICOPTER	
DATE	TRACER	START (h)	END (h)	RATE (g/s)	START (h)	END (h)	N° OF PERIODS (30 min each)	START (h)	END (h)	START (h)	END (h)
12.7.88	C ₇ F ₁₄	12.00	13.00	0.44	12.00	15.00 ⁽¹⁾	6	13.17	14.52		
	SF ₆	12.00	13.00	4.19	12.00	15.00 ⁽¹⁾	6				
19.7.88	C ₇ F ₁₄	12.00	13.00	0.54	12.00	15.00 ⁽¹⁾	6	13.16	14.44		
21.7.88	C ₇ F ₁₄	12.30	13.30	0.54	12.30	15.30 ⁽¹⁾	6	14.00	15.40		

(1) The samplers in the two arcs nearest to the release point stopped sampling one hour before.

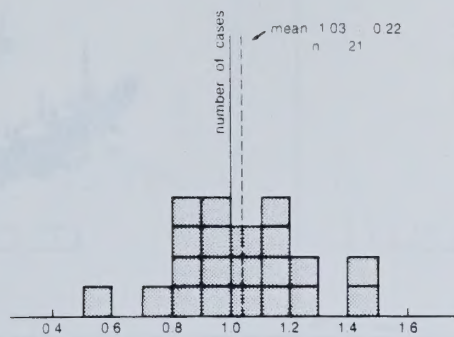


Fig. 1 Normalized SF₆/C.F. concentration ratios (averages over 1/2 hour)

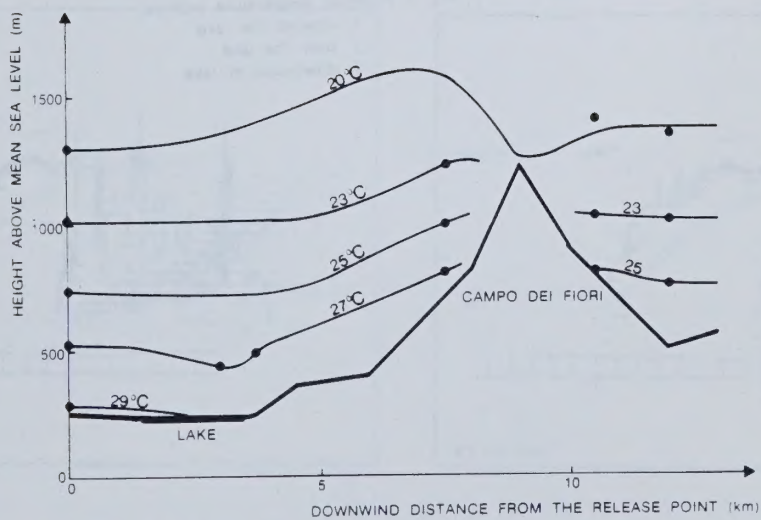


Fig. 2 Temperature isolines during the 21st July experiment

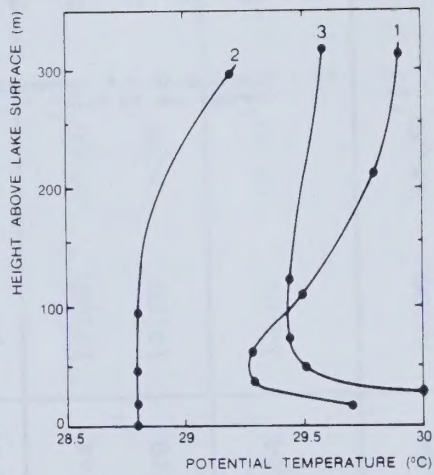


Fig 3 Potential temperature profiles:
 1 upwind the lake
 2 over the lake
 3 downwind th lake

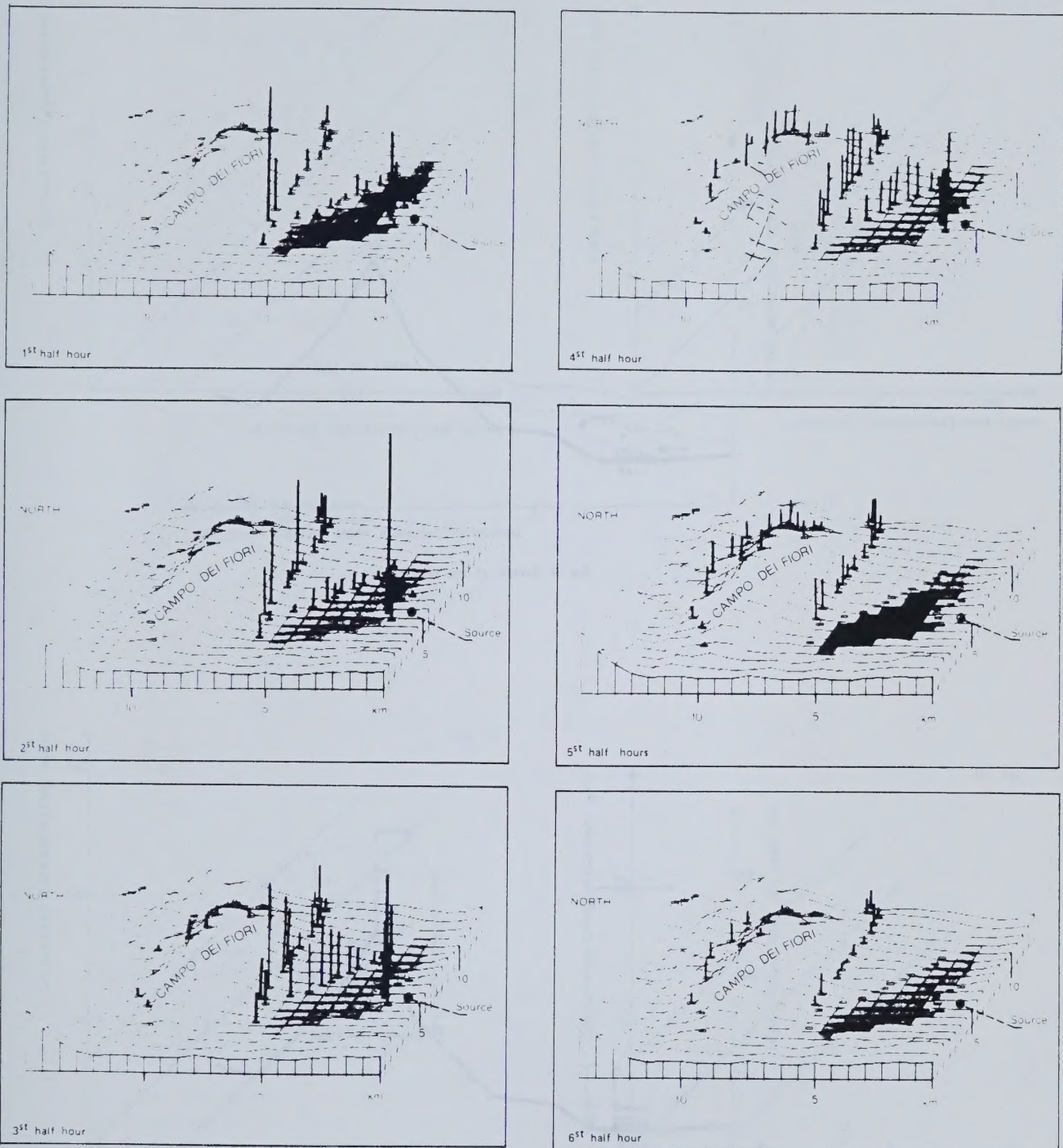


Fig. 4 Sampling network and observed tracer concentrations

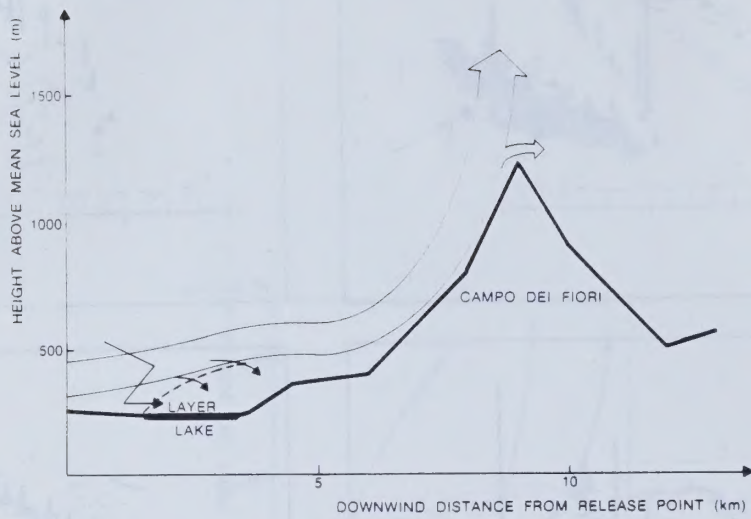


Fig. 5 Sketch of the flow field

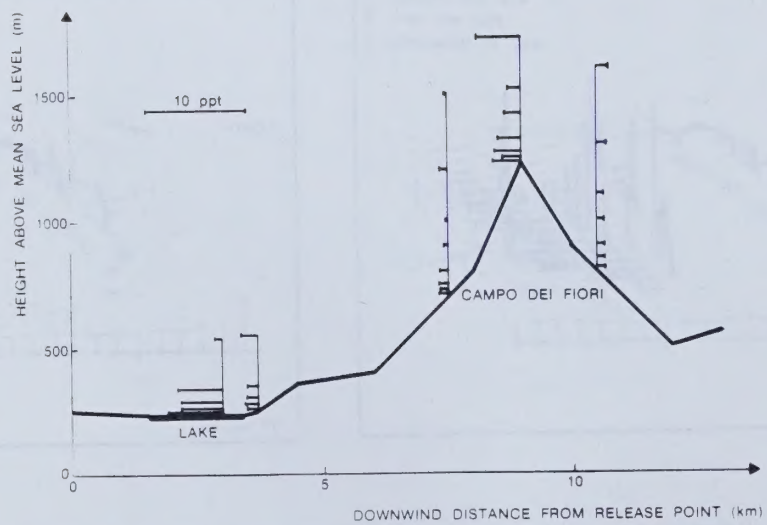


Fig. 6 Tracer concentrations measured by helicopter

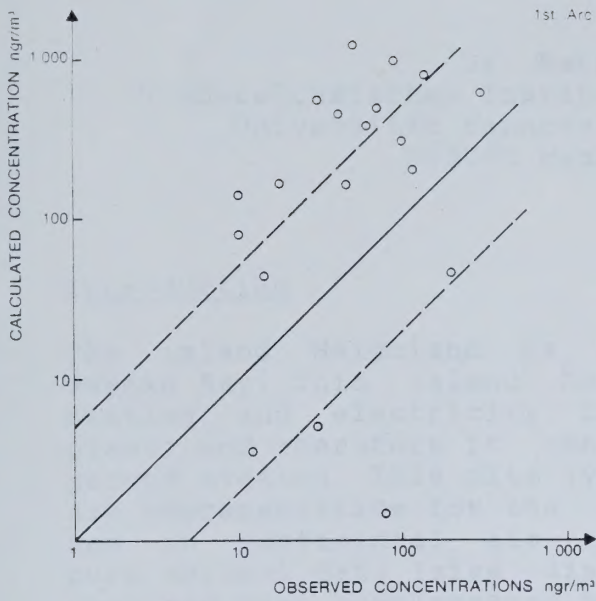


Fig. 7a The 21st July 1988
experiment simulation

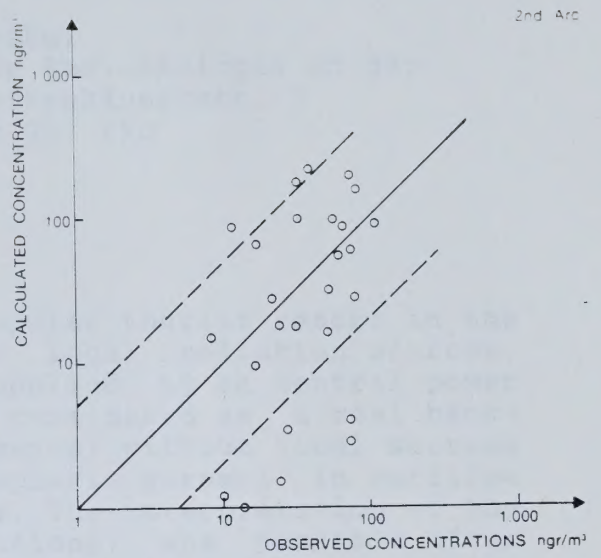


Fig. 7b

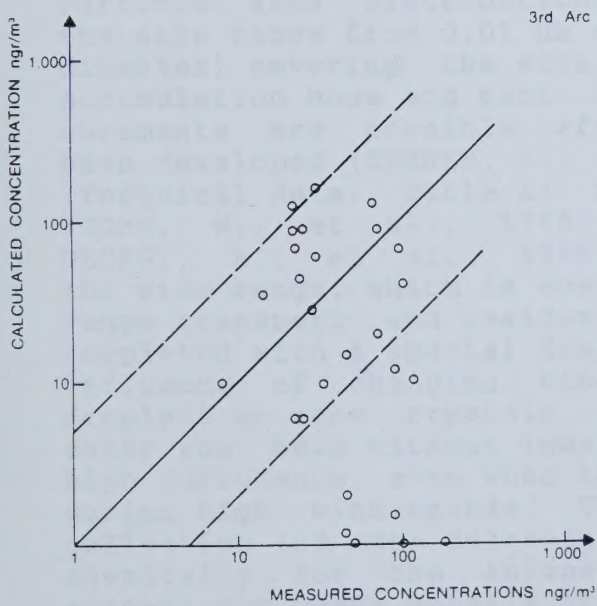


Fig. 7c

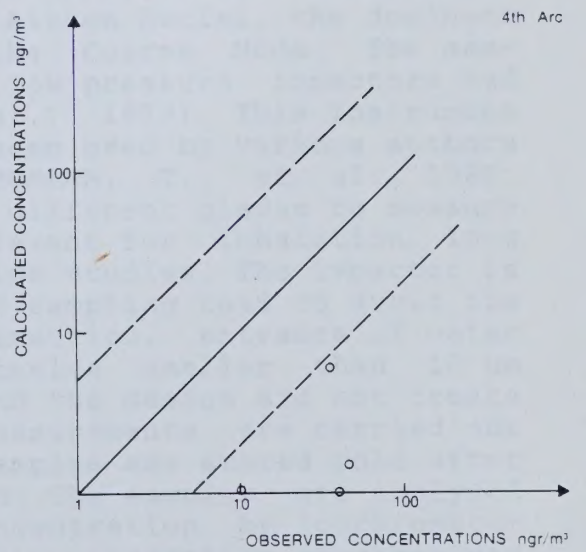


Fig. 7d



Fig. 1. Cross-section of the river.



Fig. 2. Cross-section of the river.

SIZE DISTRIBUTION MEASUREMENTS OF THE ATMOSPHERIC AEROSOL IN HELGOLAND

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Introduction

The island Helgoland is a popular tourist resort in the German Bay. This island has no local pollution sources, heating and electricity is supplied by an central power plant, and therefore it can be considered as a real background station. This site (Vogelwarte) without local sources are representative for the atmospheric aerosol in maritime and in continental air masses. The interpretation of the pure aerosol data (size distributions) was possible after they had been complemented by meteorological data, trajectory analysis, gas measurements and the concentrations of the assorted chemical constituents in the aerosol samples supplied by the LÜN (Niedersächsisches Luftüberwachungs Netz). The collected data are selected on the basis of several meteorological situations as an example of the influence and variation of airborne particulates.

Materials and Methods

Particle size distribution measurements are performed in the size range from 0,01 μm up to 15 μm d.ae. (aerodynamic diameter) covering the size of Aitken Nuclei, the dominant Accumulation Mode and part of the Coarse Mode. The measurements are possible after low pressure impactors had been developed (BERNER, A., et al., 1979). This instrument (Technical data, Table 1) has been used by various authors (JOHN, W., et al., 1986; SCHUMANN, T., et al., 1986; GEORGI, B., et al., 1986) at different places to measure the size range, which is most relevant for inhalation, long range transport and residence time studies. The impactor is completed with a special designed sampling head to avoid the influence of changing wind direction, entrance of water droplets or snow crystals. Particles smaller than 10 μm enter the head without losses and the design did not create high turbulence, even when the measurements are carried out during high wind speeds. The samples are stored cold after collection and mass determination. The samples are analysed chemically for the anions concentration by ionchromatography, for heavy metals by atomic absorption spectroscopy or for rare earth elements by neutrone activation analysis. The analytical detection limits for the atomic absorption (Pb and Cd) and the ion-chromatography (Cl^- , NO_3^- , SO_4^{2-}) measurements due to the background of the sampling foils and an average measuring volume of 100 m³ are listed in table 2.

IMPACTOR LPI 25/0,015 OPERATED WITH CRITICAL NOZZLE

Volumetric flow rate: 25 l/min (nominal value at 20 °C)

Measuring range: 0,015 µm ae.d. to 16 µm ae.d. (ae.d.: aerodynamic equivalent diameter).

Cut off diameters $D(0,1)$ µm ae.d. of $\{hg\}$ stages and average particle diameters $D(i) = (D(0,1) \cdot D(0,1+i))^{1/2}$ µm ae.d.;

i.: stage index

i	-1	0	1	2	3	4	5	6	7	8	9
$D(0,1)$	0,015	0,030	0,060	0,125	0,25	0,50	1,0	2,0	4,0	8,0	16
$D(i)$	0,021	0,042	0,087	0,18	0,35	0,71	1,4	2,8	5,7	11,3	--

Vacuum pump: minimal capacity 35 m³/h at a pressure of 40 mbar near the end flange of the impactor.

IMPACTOR LPI 30/0,06 OPERATED WITH CRITICAL NOZZLE

Volumetric flow range: 30 l/min (nominal value at 20°C)

Measuring range: 0,06 µm ae.d. to 16 µm ae.d. (ae.d.: aerodynamic equivalent diameter).

Cut off diameters $D(0,1)$ µm ae.d. of $\{hg\}$ stages and average particle diameters $D(i) = (D(0,1) \cdot D(0,1+i))^{1/2}$ µm ae.d.;

i.: stage index

i	1	2	3	4	5	6	7	8	9
$D(0,1)$	0,06	0,125	0,25	0,5	1,0	2,0	4,0	8,0	16
$D(i)$	0,087	0,18	0,35	0,71	1,4	2,8	5,7	11,3	--

Vacuum pump: minimal capacity 12 m³/h at a pressure of 150 mbar near the end flange of the impactor.

Table 1. Technical data for the low pressure impactors LPI 30 and LPI 25

Chemical constituent of the aerosol particles	absolut detection limit	concentration limit
---	-------------------------	---------------------

Pb	30 ng	≤ 0,30 ng m ⁻³
Cd	6 ng	≤ 0,06 ng m ⁻³
Cl ⁻	2,7 µg	≤ 27 ng m ⁻³
NO ₃ ⁻	1,5 µg	≤ 15 ng m ⁻³
SO ₄ ²⁻	5 µg	≤ 50 ng m ⁻³

Table 2. Detection limits for atomic absorption and ion-chromatography measurements on impactor folies and for a typical air sample of 100 m³

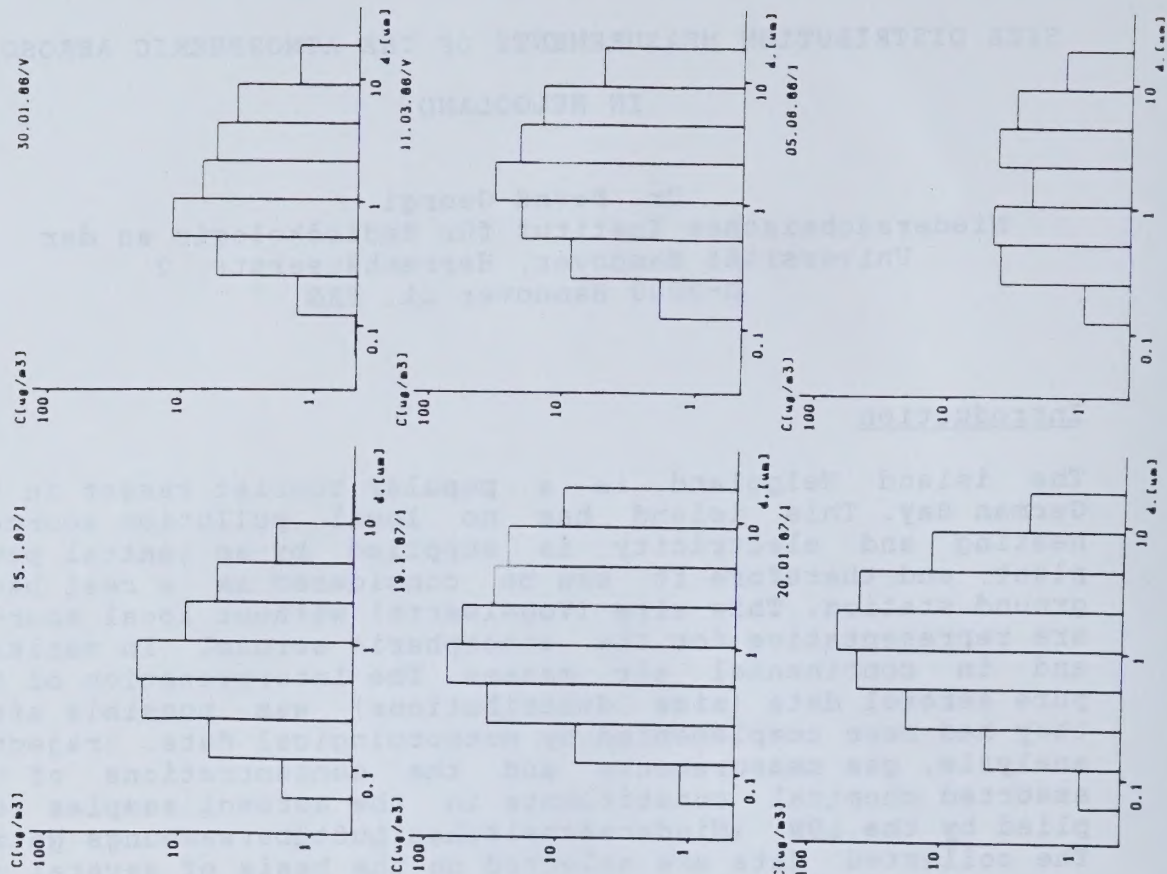


Figure 1. The comparison of three significant size distribution measurements in Helgoland (30.01., 11.03. and 05.06.1986) and Hannover (15.01., 20.01.1987)

Results

The size distribution measurements started in 1985 at Helgoland. A number of measurements are performed at some other places in Lower Saxony too. From this data base some results are selected to show characteristic and important particle concentrations in Helgoland and Lower Saxony₃. The total suspended matter (TSP) is fairly low 30-50 $\mu\text{g m}^{-3}$, if maritime air masses are present and local sources are negligible. The values increase by a factor of 5, if continental air masses and local sources are present. These two different situations are connected with distinct size distributions. The bi-modal distribution with low concentrations, a first mode around 0,5 $\mu\text{m d.ae.}$ and a second between 2 and 4 $\mu\text{m d.ae.}$ originated by sea spray. The distributions for the high concentrations are characterized by a single mode between 0,5 and 1,0 $\mu\text{m d.ae.}$ Figure 1. shows the high concentration situation for Hannover and three situations for Helgoland, which show the influence of long range transport even at the background station.

The chemical constituents are listed in table 3. The anions like Cl^- are characteristic for sea spray if the maximum occurs between 2 and 4 $\mu\text{m d.ae.}$ The second mode below 1 $\mu\text{m d.ae.}$ are significant for anthropogenic source. The single Sulfat (SO_4^{--}) mode below 1 $\mu\text{m d.ae.}$ show the influence of combustion⁴ processes. A similar finger print for the influence of automobil exhaust are the single lead mode and the amount of nitrats (NO_3^-). The occurrence of the heavy metalls especially Cd are significant for industrial activities.

CONCENTRATIONS (ng m^{-3})																
Stufe	ϕ μm	31.01.86 (289,2 m^3)					11.03.86 (342,9 m^3)					05.06.86 (288,9 m^3)				
		Pb	Cd	Cl^-	NO_3^-	SO_4^{--}	Pb	Cd	Cl^-	NO_3^-	SO_4^{--}	Pb	Cd	Cl^-	NO_3^-	SO_4^{--}
1	0,06	0,41	-	35	-	45	0,47	-	9	-	67	0,62	-	13	-	119
2	0,12	1,66	-	9	-	256	3,44	-	9	192	334	2,28	-	12	141	316
3	0,25	4,36	0,09	25	557	1478	11,2	0,22	62	1135	1680	3,53	-	33	856	1131
4	0,50	6,36	0,15	38	973	4259	18,4	0,24	283	4366	5532	3,46	0,06	55	1047	997
5	1,00	4,43	0,11	120	528	2663	24,9	0,35	502	6955	11024	3,46	-	229	1390	713
6	2,00	1,38	-	564	322	663	13,4	0,19	455	4030	6223	1,73	-	832	1911	515
7	4,00	0,35	-	639	-	224	6,8	0,09	382	2704	3944	0,76	-	668	834	292
8	8,00	-	-	194	-	46	2,3	-	136	813	1357	0,14	-	162	171	78
Σ		19	0,35	1623	2380	9634	80,9	1,09	1837	20195	30160	16		2004	6350	4161
T S P		$\Sigma = 37,3 \mu\text{gm}^{-3}$					$\Sigma = 105,1 \mu\text{gm}^{-3}$					$\Sigma = 21,4 \mu\text{gm}^{-3}$				

Table 3. Chemical constituents of three aerosol measurements in Helgoland.

Discussion

The measured size distributions and the analysed chemical constituents combined with the meteorological data show the influence of long range transported air pollutants as main part of the immission burden. The consequence of the high concentrations (up to 60%) of particle $\leq 1 \mu\text{m d.ae.}$ are a long residence time, the domination of wet removal processes and possible adverse health effects due to their inhalation.

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Summary and Conclusions

The following conclusions were drawn from the study:

The study showed that the majority of the respondents were male, aged between 20 and 30 years, and had a high level of education. The majority of the respondents were employed in the private sector, and the majority of the respondents were from the urban areas.

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Summary

The study showed that the majority of the respondents were male, aged between 20 and 30 years, and had a high level of education. The majority of the respondents were employed in the private sector, and the majority of the respondents were from the urban areas. The study also showed that the majority of the respondents were from the urban areas, and the majority of the respondents were from the urban areas. The study also showed that the majority of the respondents were from the urban areas, and the majority of the respondents were from the urban areas. The study also showed that the majority of the respondents were from the urban areas, and the majority of the respondents were from the urban areas.

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Summary and Conclusions

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The views expressed below are entirely my own and do in no way reflect the consensus of the participants of the meeting. I have divided my comments into three groups: 1. Papers related to the Öresund experiment, 2. Studies from other coastal areas and 3. Concluding remarks.

1. The Öresund experiment

A. The main problems:

· Characteristics of the flow field

Phenomenological models:

Melas has presented a quite successful model for the development of the stable internal boundary layer that develops during daytime over Öresund; *Gryning et al* have made corresponding, successful modelling work related to the development of the convective IBL that evolves over the downwind coast.

Numerical models:

Gryning and Doran, *Uliasz* and *Andre'n* have presented simulations with different models. These simulations show qualitative (and sometimes quantitative) agreement with the measurements and reveal several interesting physical mechanisms.

· The dispersion experiments: Several attempts to simulate the observed concentration distributions in the seven tracer experiments have been made. These attempts either fail to simulate the results of the experiment on June 5 (typically calculated concentrations are wrong by a factor 2 - 4) or the concentration fields of all the other experiments. It is striking that even a very sophisticated mesoscale model (*Andre'n*) fails to solve this problem. It is probable that some physical mechanism that is not adequately treated in the present models is responsible for the failure, maybe synoptic scale processes, but there are other possibilities as well as indicated in the discussions during the meeting.

B. Other studies that have used Öresund data but concentrate on other aspects

Smedman's work on the structure of stable boundary layers indicate the usefulness and the high quality of the Öresund data set for more general studies of relevance to boundary-layer meteorology.

2. Studies from other coastal areas

A. Areas without high mountains

Phenomenological models: *Bergström* has successfully shown that it is possible to treat the relevant problems of the combined effect of a change in roughness and of low hills on the turbulence structure. *Sempreviva et al* have illustrated how the growth of an internal boundary layer in neutral conditions can be modelled with relatively simple methods.

B. Areas with high mountains

Several good illustrations were given - with model simulations and, to a lesser extent, with measurements - of complicated combinations of sea breezes, anabatic winds and topographical channeling of the flow, including the effect on dispersion from point sources: *Barzis*, *Physick*, *Uliasz*, *Gaglione et al* and *Glerici et al*.

3. Concluding remarks

·The results of the Öresund data point towards the limitations in our basic understanding of the interplay between different physical factors even in a relatively simple topographical situation and in our ability to model it.

·Most likely the situation is even worse when complicated terrain is encountered, although there may be fortuitous factors that simplify the situation, like strong local topographic forcing, cf. the very promising wind field verification presented by *Physick*. This result is in general agreement with findings in Sweden, where it was found that model predictions agree quite well with measurements in an area with modest topographical relief (ca. 80m).

·There is a need for good field experiments, where all relevant parameters are measured. In particular there is the need of high quality data with good spatial resolution well above the surface, i.e. aircraft or helicopter data.

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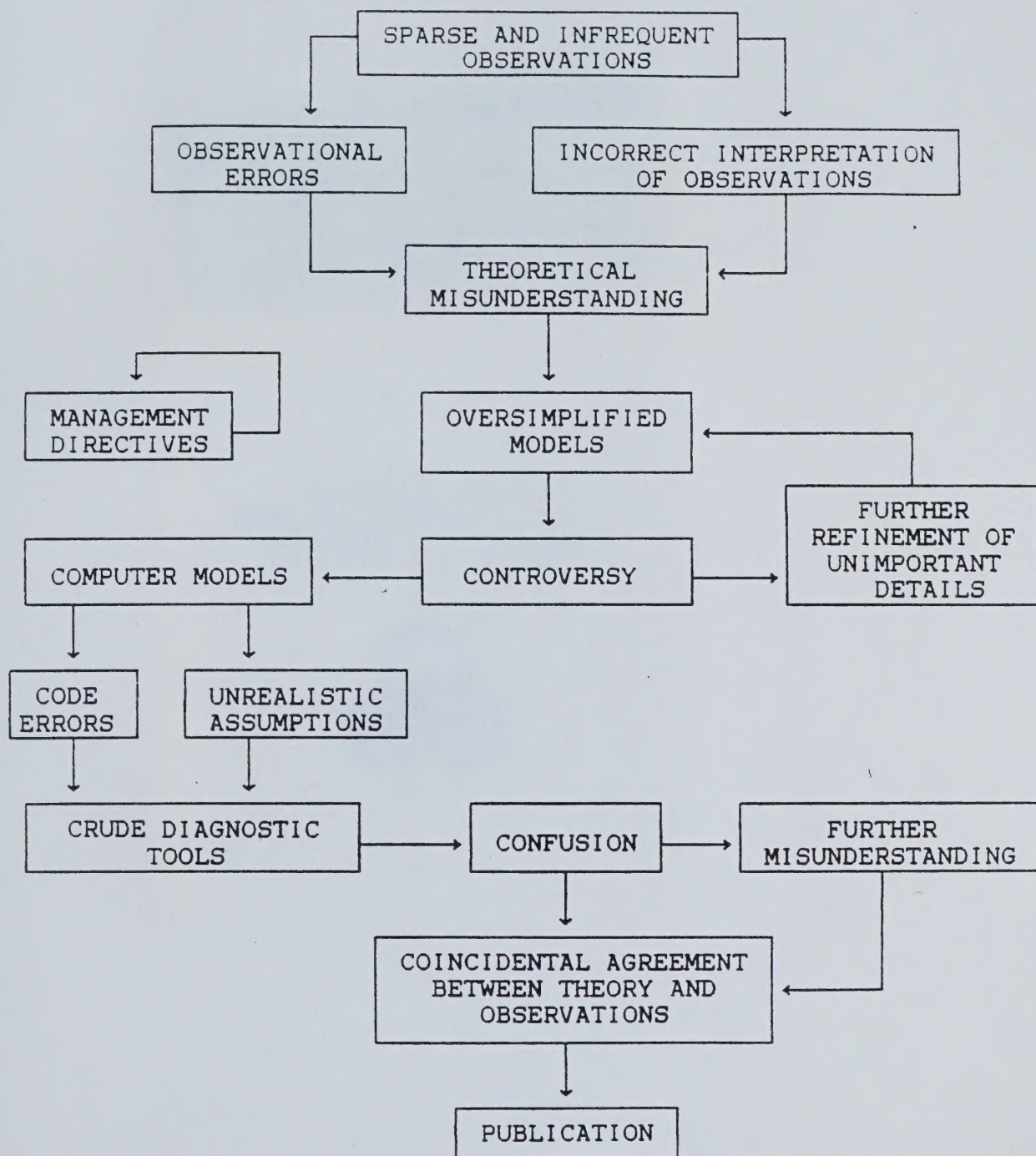
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